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# The Torridonian Succession of the Isle of Rhum

By G. P. BLACK and W. WELSH

ABSTRACT

The Torridonian rocks of Rhum are divided into five sub-divisions on lithological grounds and the succession is found to be closely comparable with that of Skye. On the basis of the comparison, four sub-divisions, totalling 8,500 feet in thickness, are allocated to the Diabaig Group while the 6,000 feet of strata belonging to the fifth and highest sub-division are allocated to the Applecross Group. The sediments of the Diabaig Group are now seen to form a great wedge thickening and coarsening consistently from Gairloch south-south-westwards to Rhum ; a phenomenon obscured by the previously held correlation of the Rhum Torridonian with that of Skye and the Mainland.

## I. INTRODUCTION

THE Isle of Rhum lies seven miles south of Skye and fifteen miles north of Ardnamurchan Point. Torridonian rocks underlie some 20 square miles of the island and this tract forms the most southerly major occurrence of the Torridonian System to the north-west of the Great Glen. The emplacement of the Tertiary Igneous Complex of Rhum took place partly at the expense of the Torridonian, but the associated tectonic disturbances were confined to the immediate neighbourhood of the igneous masses and the greater part of the Torridonian outcrop remained essentially undisturbed.

Over 14,000 feet of Torridonian strata are present on Rhum, resting unconformably on Lewisian Gneiss. The strata dip to the north-west at angles varying from 10° to 70°, the general dip being 20° to 25°. Excellent exposures are to be found in the almost continuous cliffs which mark the seaward limit of the Torridonian outcrop ; inland many exposures occur, but do not accurately portray the relative abundances of the various rock types, the coarser beds cropping out more freely than their finer associates.

The earliest significant description of the Torridonian rocks of Rhum is that by Harker (1908) who divided the sequence into two units, a lower shale group 1,400 feet thick being surmounted by 9,000 feet of sandstone. In the sandstone group he noted the rarity of shaly beds, the occurrence of current bedding and the existence of slump structures—" violent contortions in beds which are divided by regular parallel bedding planes " (ibid., p. 6). Harker considered that both autochthonous and exotic Torridonian rocks were present and much

of his description was devoted to the effects of the supposed thrusting. He did not attempt to correlate the Torridonian of Rhum with that of Skye or of the mainland.

In 1945 Bailey revised in some measure Harker's work on Rhum. He concluded that no exotic Torridonian rocks occurred on Rhum and re-identified Harker's thrust as a ring fault of Tertiary age. The existence of Lewisian Gneiss on the island was recognized, and Bailey pointed out that in several localities the Gneiss was unconformably covered by a Torridonian basal grit. His division of the Rhum Torridonian was thus tripartite, at least 200 feet of basal grit being succeeded upwards by the shale group which in turn was overlain by the sandstone group. Bailey correlated the basal grit and the succeeding shales with the Diabaig Group, and the overlying sandstones with the Applecross Group, of Skye.

Since 1945, no work devoted primarily, or even extensively, to the Rhum Torridonian has been published. Bailey's correlation of the Torridonian of Rhum and Skye has been accepted by later writers (e.g. Phemister, 1948, Hughes, 1960) and reaffirmed by Bailey in a further note on Rhum (1956). Some authors, however, have shown caution in its acceptance (Richey, 1948).

## 2. The Torridonian Succession

The present investigation has shown that the upper sandstone group of Harker and Bailey consists of at least three distinct units. To avoid confusion, and to facilitate discussion of the probable correlation of the Rhum and Skye successions, local names have been given to the various subdivisions on Rhum.

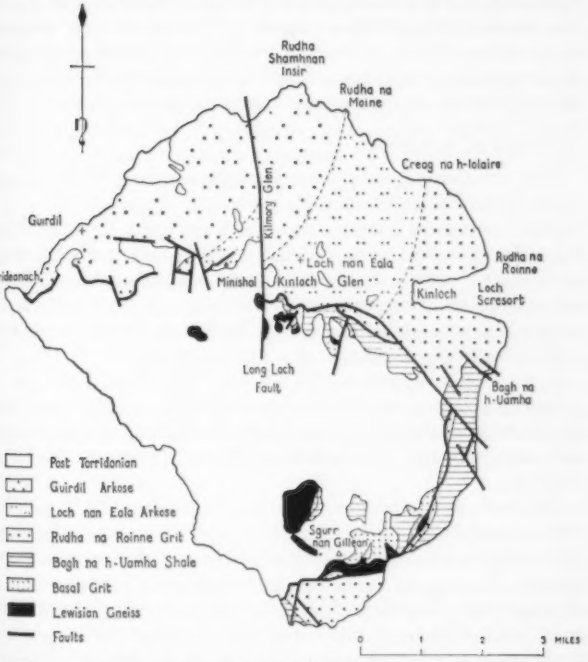
### (*a*) *Basal Grit*

Bailey (1945, p. 170) tersely described the rocks of this unit as pebbly grits containing fragments of the underlying Lewisian Gneiss. He assigned to the Basal Grit outcrops on the slopes of Sgurr nan Gillean in the south of Rhum which Harker (1908, p. 13) had described as arenaceous beds containing numerous fragments of gneiss.

The Basal Grit rocks have been found to be highly variable in character, the most typical variety being a poorly sorted grit consisting of angular and sub-angular fragments of quartz and feldspar set in a matrix of secondary quartz. Feldspar is relatively abundant, common varieties being plagioclase, microcline, perthite and microcline-perthite; orthoclase is only sparsely present. Chloritic alteration is common. In some specimens abundant grains of epidote are present, probably at least in part authigenic, although the presence of sparse epidosite fragments suggests that some of the epidote may be detrital. Even in the coarsest specimen examined in thin-section rock fragments

are rare, large pebbles consisting of a single mineral. Fragments which have been identified include vein quartz, epidosite and rhyolite.

The Basal Grit of Rhum thus resembles the Epidotic Grits and Conglomerates found at the base of the Torridonian sequence in Skye (Clough, 1907, pp. 353–354). The latter have been much affected by Caledonian crustal movements whereas the former outcrops in a region of intense igneous activity and has largely suffered Tertiary thermal




TEXT-FIG. 1.—The distribution of the sub-divisions of the Torridonian in Rhum.

metamorphism with local cataclasis. In both the Skye and Rhum occurrences, therefore, the original constitution and texture of the rocks have been modified, but not sufficiently to obscure the resemblances between them.

### (*b*) *Bagh na h-Uamha Shale*

This sub-division, equivalent to the shale group of Harker and Bailey, is well-exposed along the eastern shores. Some 1,400 feet in

thickness, it consists of dark coloured shales and siltstones showing more or less marked lamination. Towards the top it contains intercalated beds of white sandstone of varying thickness, and appears to pass upwards into the overlying sub-division by increase of the proportion of arenaceous material.

Thin-section examination of specimens collected high in this sub-division show the rocks to be siltstones of varying granularity consisting essentially of sub-angular quartz grains set in an argillaceous or very fine quartz-chlorite matrix. Feldspar (orthoclase) is rare. Occasional grains of zircon and epidote are present. A specimen from a lower horizon proved to be a shale consisting of sparse angular quartz grains set in a very fine-grained matrix of quartz and chlorite.

(*c*) *Rudha na Roinne Grit*

This sub-division corresponds to the lowest part of the sandstone group of Harker and Bailey. It is exposed in the east around Loch Scresort, occupying a tract lying to the north and west of the Bagh na h-Uamha Shale and east of a line trending north-north-east through Kinloch and reaching the coast at Creag na h-Iolaire. The thickness of the sub-division is approximately 4,000 feet. Its boundaries are well marked but transitional against the Bagh na h-Uamha Shale below, and less well-defined against the succeeding rocks.

In the field the rocks of this sub-division have a characteristic pinkish-grey colour, seen in the coarsest beds to be due to their content of pink feldspar and grey quartz. They consist of alternations of pebbly grits with beds of finer arkosic sandstone. Current-bedding is very marked.

Thin-section examination of a series of specimens collected from the cliffs around Rudha na Roinne shows the rocks to have a very simple mineralogy, consisting almost entirely of quartz (or vein quartzite) and feldspar. In general the grains are sub-angular and poorly sorted although there is better sorting in the finer members. Microcline and orthoclase are abundant in all the sections examined, but the proportion of plagioclase varies, in some specimens decreasing to zero. In the finer specimens plagioclase is more prominent than in the coarser, but it is nowhere as abundant as microcline. In composition the plagioclase is highly sodic, approaching pure albite. Microcline-perthite and plagioclase-antiperthite occur in small amount. Heavy minerals are present in very small proportion, only one or two grains being observed in a typical thin-section. Zircon strongly preponderates; muscovite and rare biotite also occur.

Even in the coarsest rocks examined, rock fragments, other than vein quartz, occur but sparsely. Despite their small proportion they include a large variety of rock types—chert, metamorphic quartzite,

quartz-plagioclase, quartz-plagioclase-epidote, quartz-microcline, quartz-epidote, quartz mica-schist, microcline-perthite, plagioclase-antiperthite, arkose, greywacke, fine siltstone, fine shale, felsite, and porphyrite. In the coarsest rocks the pebbles include large quartz and feldspar fragments.

The constituent grains and fragments are closely compacted so there is little matrix. Interstices between grains are generally filled with secondary silica, though rarely this is accompanied by some secondary haematite. In some specimens a proportion of argillaceous material appears in the matrix.

Secondary mineralization includes haematisation of many of the feldspar grains to give the characteristic pink colouration. A considerable proportion of the feldspar grains have been sericitized, some grains being almost entirely replaced.

### (*d*) *Loch nan Eala Arkose*

Above the Rudha na Roinne Grit lies a series of finer arkoses some 2,800 feet in thickness. These occupy the tract extending from Kinloch Glen northwards to the north-eastern coast. The tract is bounded on the east by the outcrop of the Rudha na Roinne Grit and extends westwards to a line trending from Minishal in the south to Rudha na Roinne in the north. The Long Loch Fault crosses the tract obliquely near its western margin ; the throw of the fault can be estimated as 800 feet from the displacement of the apparently concordant peridotite sill forming the northernmost tongue of the basic and ultrabasic complex of the island.

In hand-specimen the rocks of this sub-division are pale red arkoses. Ripple marking occurs at some horizons and locally slump structures are excellently developed. In thin-section the rocks are seen to be very poorly sorted, a characteristic found even in the fine-grained specimens. Sub-angular quartz and orthoclase grains are set in a matrix of fine quartz with or without chlorite. Plagioclase—albite or oligoclase—is commonly present, but always in small amount, and in some specimens a small proportion of microcline occurs. A fourth feldspar is occasionally present and shows a very small negative optic axial angle ; tentatively identified as sanidine, it occurs both as independent grains and in rock fragments.

Rock fragments are sparse, but include, in addition to the ubiquitous vein quartz, trachyte, rhyolite, felsite, greywacke, fine siltstone, chert, quartz-porphyry, quartz-orthoclase, quartz-plagioclase-muscovite, quartz-epidote, and orthoclase-epidote.

Little secondary quartz is present in the rocks except where they are altered near the main Rhum granophyre (Black, 1954), and the feldspar

TEXT-FIG. 2.—The outcrop of Torridonian strata from Rhum to Gairloch showing the localities mentioned in the text. The figures at each locality indicate the thickness there of the Diabaig Group in feet.

grains have been only slightly affected by haematisation ; alteration to secondary mica is widespread.

In comparison with the underlying Rudha na Roinne Grit, the Loch nan Eala Arkose contains a much higher proportion of the finer grain sizes and a less comprehensive suite of rock fragments. The Arkose is more poorly sorted so that an argillaceous matrix is commonly present. A further contrast is found in the relative proportions of the feldspars.

In the Rudha na Roinne Grit microcline is dominant, exceeding plagioclase in abundance ; in the Loch nan Eala Arkose microcline is not always present and is generally subordinate to the plagioclase. A further distinction lies in that the red staining due to haematisation is characteristic of the Rudha na Roinne Grit, but not of the Loch nan Eala Arkose.

(*e*) *Guirdil Arkose*

Above the Loch nan Eala Arkose lies a further series of arkosic beds, the Guirdil Arkose, the highest sub-division of the Torridonian occurring on Rhum. Rocks of this sub-division form a tract lying to the north-west of the Loch nan Eala Arkose and extending from Kilmory Glen and Rudha Shamhnan Insir to Guirdil and almost to A'Bhrideanach. The top of the sub-division is truncated by the sea and the base is ill-defined ; a total thickness of approximately 6,000 feet of strata appears to be present.

The rocks comprising the Guirdil Arkose are red to white in colour. Throughout the sub-division there appears to be a general upwards decrease in grain-size, and argillaceous beds appear in the higher part. The rocks are generally well-sorted, a feature which distinguishes them from the underlying Loch nan Eala Arkose. In thin-section they are seen to consist essentially of sub-angular quartz grains set in a matrix of secondary silica with or without chlorite. Feldspar occurs in fair proportion, the amount decreasing in the finer rocks ; orthoclase and a highly sodic plagioclase (near pure albite) are the most common varieties. Microcline also occurs but is normally subordinate to plagioclase ; a few grains of perthite have been found. In the coarser beds near the base of the sub-division, fragments of vein quartz, metamorphic quartzite, fine siltstone, fine greywacke, mica schist, rhyolite, felsite, quartz-orthoclase, quartz-epidote, and quartz-plagioclase arkose were identified. Detrital muscovite is found in the higher beds while bands rich in iron-ores occur in the lower part of the sub-division. In the lower beds zircon was found to be the only common heavy mineral, apart from ores, whereas in some beds high in the succession a much fuller suite of heavy minerals—epidote, zircon, and sphene with tourmaline, apatite, garnet, and magnetite—was discovered.

Secondary chlorite is common especially in the higher part of the sub-division and the rocks throughout show extensive haematisation of the feldspars.

Apart from the better sorting found in the rocks of this sub-division, they differ from the underlying Loch nan Eala Arkose in that they contain much secondary silica as cement, show intense haematisation of the feldspars and contain a fuller suite of heavy minerals.

### 3. Discussion and Correlation

The Torridonian succession of Rhum consists of five sub-divisions whose main characteristics are summarized below :—

5. Guirdil Arkose (6,000+ feet). Alternating coarse and fine well-sorted arkoses becoming finer in grain upwards. Feldspar not abundant; microcline subordinate.
4. Loch nan Eala Arkose (2,800 feet). Fine, poorly sorted arkose. Microcline rare or absent.
3. Rudha na Roinne Grit (4,000 feet). Well-sorted red grits and arkoses with much current bedding. Microcline the dominant feldspar.
2. Bagh na h-Uamha Shale (1,400 feet). Dark shales and siltstones with intervalated sandstones becoming prominent in the highest part.
1. Basal Grit (300 feet). Breccias, grits, and sandstones of variable lithology.

In Skye the Torridonian succession likewise comprises five separable sub-divisions (Clough, 1907, p. 349) :—

*Applecross Group.*

5. Applecross Grit (5,000+ feet). Red and chocolate arkoses.

*Diabaig Group.*

4. Kinloch Beds (3,500 feet). Sandy shales and fine grits.
3. Beinn na Seamraig Grit (2,600 feet). Fine grits with some sandy shale.
2. Loch na Dal Shale (600 to 1,200 feet). Dark shales with some fine grits.
1. Basal Grit (300 feet). Pebbly grits with epidote, sandy shales.

The strata here lie in an area of tectonic complexity and have suffered displacement in the Caledonian Orogeny.

Further to the north, on the mainland to the east of Kyle of Lochalsh, the Torridonian succession again shows a five-fold division (Peach and Horne, 1907, p. 342) :—

*Applecross Group.*

5. Applecross Group (thickness not recorded). Fine grits and sandstones.

*Diabaig Group.*

4. Kinloch Beds (2,700 feet). Sandy shales, fine sandstones, and grits.
3. Beinn na Seamraig Grit (1,200 feet). Fine grits, sandstones, and flags.
2. Loch na Dal Shale (500 feet). Dark shales with grit bands.
1. Basal Grit (60 feet). Epidotic grits with a basal conglomerate.

Still further to the north the strata of the Diabaig Group are found at Diabaig itself where the succession has been recorded by Hinxman (1907, p. 324) as :—

3. Flaggy grits and sandstones.
2. Dark shales and mudstones with thin arenaceous bands.
1. Red flaggy grits and basal breccia.

The total thickness of the Group here is some 700 feet.

From Diabaig through Lochalsh to Skye the Diabaig Group increases tenfold in thickness. From Lochalsh to Skye the increase is seen to be largely due to a thickening of the three lower members of the Group, the Kinloch Beds remaining relatively unchanged.

Bailey's correlation (1945) of the Torridonian rocks of Rhum with those of Skye and the mainland equates the entire Diabaig Group of the

latter localities with the Basal Grit and Bagh na h-Uamha Shale of Rhum. On this correlation the southward thickening of the Diabaig Group must be abruptly reversed between Sleat and Rhum so that 7,200 feet of strata in Skye are represented by only 1,700 feet in Rhum. Such a drastic southwards attenuation of the group, especially in view of consistent thickening in a southerly direction elsewhere, could only be accepted if it were based on a convincing correlation of the strata based on lithological resemblances which leave no reasonable doubt as to the stratigraphic equivalence of the strata which are being equated. Yet comparison of the Diabaig Group of Skye with Bailey's " Diabaig Group " in Rhum reveals almost no lithological similarities, but instead provides a striking contrast. The dominantly arenaceous sequence in Skye is replaced in Rhum by a succession consisting almost entirely of dark shales and siltstones overlying a relatively thin Basal Grit. There is thus an almost complete lack of lithological correspondence between the Diabaig Group of Skye and of Rhum, if Bailey's correlation is accepted. On the other hand sufficient lithological resemblances between the entire Torridonian successions of the two islands exist to justify the proposal of a new correlation.

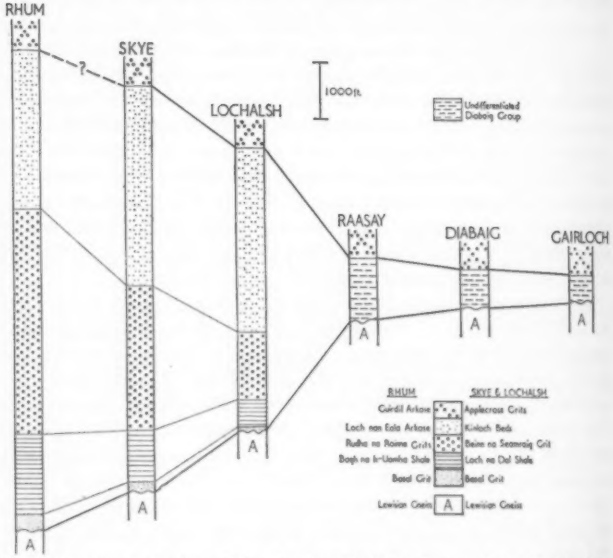
The Bagh na h-Uamha Shale of Rhum is closely similar to the Loch na Dal Shale of Skye. Both consist of dark shales or siltstones with intercalated beds of sandstone. They are of similar thickness, the Rhum sub-division being slightly the thicker, a phenomenon in accordance with the general southwards thickening of the Loch na Dal Shale from Lochalsh to Sleat. Both sub-divisions lie on a grit and both are covered by grits ; both occupy a similar stratigraphic position, their bases lying a few hundred feet above the Torridonian/Lewisian unconformity. In the authors' opinion there is no doubt that the Bagh na h-Uamha Shale is the Rhum equivalent of the Loch na Dal Shale.

The correlation of the two shale units leads immediately to a correlation of the Basal Grit of Rhum with that of Skye and Lochalsh. These have a similar thickness and a similar, if variable, lithology ; both are characterized by the local occurrence of abundant epidote. Their stratigraphic positions are identical, intervening between the Lewisian Gneiss and the overlying shale horizon.

In Skye and Lochalsh the Loch na Dal Shale is covered by the Beinn na Seamraig Grit, a sub-division of the Torridonian increasing in thickness from 1,200 feet in the north to 2,700 feet in the south. In Rhum the Bagh na h-Uamha Shale is covered by the 4,000 feet of the Rudha na Roinne Grit, a sub-division consisting of grits and arkoses similar to, but apparently rather coarser than, the fine grits with some sandy shale which form the Beinn na Seamraig Grit of Skye. Stratigraphic position and lithological resemblance combine to suggest that

the Rudha na Roinne Grit of Rhum is equivalent to the Beinn na Seamraig Grit of Skye and Lochalsh.

Above the Rudha na Roinne Grit, there occurs a series of arkoses divisible into two sub-divisions—the Loch nan Eala and Guirdil Arkoses—on differences in sorting, secondary haematisation of feldspars, heavy mineral content, and, to some extent, grain-size. The boundary between the two sub-divisions is, however, transitional and




TEXT-FIG. 3.—Lateral variation in the Diabaig Group from Gairloch to Rhum.

poorly defined although the sub-divisions themselves are distinct. In Skye a dominantly arenaceous series lies above the Beinn na Seamraig Grit and has been divided into two sub-divisions (Clough, 1907) along a boundary which, in part of Sleat at least, " represents a boundary between different types of grit " (p. 358). Clough, however, gives no criteria for distinguishing between his two types of grit and it therefore cannot be definitely assumed that the boundary between the Kinloch Beds and the Applecross Grit in Skye is at the precise stratigraphic horizon of the boundary between the Loch nan Eala and Guirdil Arkoses in Rhum. It can be said, however, that the Loch nan Eala Arkose of Rhum is the approximate equivalent of the Kinloch Beds and the Guirdil Arkose the approximate equivalent of the Applecross Grit.

On the basis of this new correlation, it at once appears that the Diabaig Group of Rhum is 8,500 feet in thickness in comparison to 7,200 feet in Skye, 4,500 feet in Lochalsh and 700 feet at Diabaig. The southerly thickening of the group as a whole, so well-marked from Diabaig to Skye, therefore continues further to the south. The major part of the increase in thickness between Skye and Rhum is due to great proportional increases in two sub-divisions—the Bagh na h-Uamha Shale and the Rudha na Roinne Grit. This corresponds closely to the change in the Diabaig Group between Lochalsh and Skye, where the two equivalent sub-divisions—the Loch na Dal Shale and the Beinn na Seamraig Grit—also show the greatest proportional increase in thickness.

Associated with this southerly increase in thickness there occurs a southerly increase in grain-size. Whereas in Skye, above the Loch na Dal Shale, argillaceous beds occur as an important although subsidiary, part of the dominantly arenaceous sequence of the Diabaig Group, in Rhum the equivalent sub-divisions contain little argillaceous material. The arenaceous beds, furthermore, appear to be coarser in Rhum than in Skye, and this coarsening of the higher part of the Diabaig Group, like the increase in thickness, appears to be a continuation of a change found between Skye and Diabaig village. In the north, at Diabaig and Raasay (Hinxman, 1907, p. 340), the Diabaig Group contains a relatively high proportion of fine-grained rocks ; this proportion decreases southwards through Lochalsh to Skye and, as shown above, continues beyond Skye, further southwards, to Rhum.

From the southward increase in thickness of the Diabaig Group as a whole and from its general tendency to become coarser in grain in the same direction, it is concluded that the source of detritus which supplied the sediments lay to the south or south-west.

Further studies of the direction of the detritus-supplying palaeocurrents are required to amplify this conclusion but a similar source has been suggested by Sutton and Watson (1960) for the materials of the Torridonian Basal Grit of Skye.

The Guirdil Arkose cannot be directly compared in thickness with its equivalents in Skye and Lochalsh as the original thicknesses are unknown. In granularity it appears to resemble the Applecross Grit of Sleat and Lochalsh rather than that of Raasay or Applecross itself, in that it is relatively fine in grain. There is thus no southerly increase in granularity in the Applecross Group but rather an ill-defined decrease ; this contrast between the Diabaig and Applecross Groups perhaps indicates that their sediments were derived from different sources, a conclusion supported in Rhum at least by their differing suites of heavy minerals.

## IV. Acknowledgments

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# The Mid-Zambezi-Luangwa Rift and Related Carbonatite Activity

By D. K. Bailey

Abstract

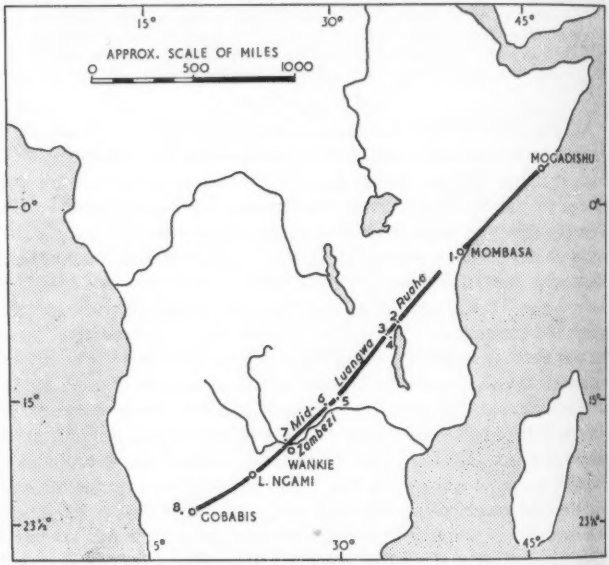
The known carbonatite eruptive centres in Northern Rhodesia fall within a linear belt trending N.E.–S.W. the most conspicuous elements of which are the colinear troughs of the mid-Zambezi valley and the Luangwa rift. This lineament is an important and long-standing zone of crustal weakness. Its extensions outside Northern Rhodesia are briefly indicated and its ramifications inside the Territory are examined, particularly in reference to the localization of carbonatite activity around rift intersections.

## I. Introduction

An important, linear structural element of Northern Rhodesia is marked by the colinear, N.E.–S.W. troughs of the mid-Zambezi valley, from Walker's Drift to Chirundu, and the Luangwa valley, from Beit Bridge to the headwaters near the Tanganyika border (Text-fig. 2). This has for long been referred to as the mid-Zambezi-Luangwa rift, although rift-faulting appears to be poorly developed on the Southern Rhodesian side of the mid-Zambezi valley. Terrestrial sediments of Karroo age (U. Carboniferous—early Jurassic) underlie these troughs which are presumed to have been formed during immediately post-Karroo earth-movements, being subsequently infilled by Cretaceous sediments and re-excavated by later cycles of erosion (Dixey, 1955). It would appear that this was a zone of marked crustal weakness in Karroo and post-Karroo times, and it may be inferred, as with most other rift zones, that it was the site of earlier instability and probably a locus of movement since the Pre-Cambrian. In any event it is clear that the mid-Zambezi-Luangwa rift, which itself stretches for 750 miles, is a segment of a much longer crustal lineament (Text-fig. 1) which stretches north-east along the colinear Ruaha rift, in Tanganyika, and may well carry through to the long, steeply shelving, straight Somalia coastline from Mombasa to Mogadishu, and beyond. Although this segment of the African coastline appears to have been stable in the Tertiary and post-Tertiary it is a remarkable feature and may, as suggested by Dixey (1956, p. 35) be following an older structural line.

In the other direction, south-west from the mid-Zambezi trough, the tectonic line is continued through the important Deka-Wankie faults, in Southern Rhodesia to the Lake Ngami-Gobabis tectonic line, running south-west through Bechuanaland into South-West Africa (Du Toit, 1954, p. 572). Wayland (1956) suggests the presence of filled-in rift-valleys along this line in Bechuanaland and South-West Africa. Throughout most of its length in the south-westerly part of the belt, from the mid-Zambezi to Gobabis, the direction closely parallels the structures in the older rocks.

This 3,000 mile linear feature may in places be a true rift valley, in others a zone of faulting and flexuring ; inflexions in its trend are produced by accommodation to structural trends in the basement, as in the south-west, or by intersection with other rift zones, more recently active, as in Tanganyika. Differences in present day expression of the feature may also result from different ages of movement predominating in different sections of its length.




TEXT-FIG.1.—The extent of the mid-Zambezi-Luangwa lineament. The numbered points are as follows : (1) Mrima ; (2) Panda Hill and Rungwe ; (3) Nkumbwa Hill ; (4) Ilomba and Mivula ; (5) Rufunsa ; (6) Keshya and Mkwisi ; (7) Livingstone ; (8) Auas mountains, Windhoek-Rehoboth.

Several important eruptive centres lie on this lineament ; the Mrima carbonatite in the Kenya coastal belt may be an example ; the Panda Hill carbonatite and the Rungwe volcanic centre lie close to the intersection of the Ruaha-Luangwa rift and the Rukwa-Nyasa rift. In Northern Rhodesia the Nkumbwa Hill carbonatite in the upper Luangwa valley (Reeve and Deans, 1954) and the Rufunsa carbonatites (Bailey, 1960) in the lower Luangwa valley are in the same belt ; the Keshya and Mkwisi carbonatites (Bailey, in press) are similarly disposed, and the marked development of the Batoka basalts of Stormberg age, around Livingstone, may be a further indication of

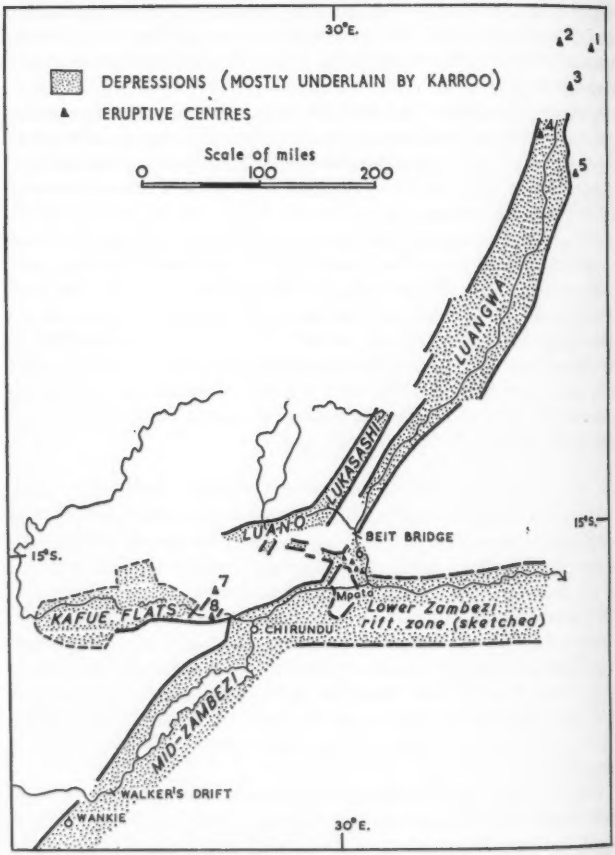
crustal instability along this zone. The extension south-west along the Wanki-Ngami-Gobabis line terminates in the alkaline volcanic area between Windhoek and Rehoboth in South-West Africa (Rennie, 1927 ; Gevers, 1934).

It should be stressed that it is the zone of structural weakness, not the disposition of the eruptive centres, that defines the lineament ; the converse of this, proposing a lineament based on the distribution of igneous activity is liable to invalidation as soon as new centres of activity are discovered, as shown by King and Sutherland (1960, p. 302). Where, however, igneous complexes of different ages lie along an already known belt of long-established crustal instability they serve only to confirm its reality, and it is appropriate to inquire into their relationships. It is quite clear that there are lineaments other than the one under discussion that are equally, or even more conspicuous in South and East Africa, but it is not appropriate for this writer to inquire into their related igneous activity : it is the relationship of carbonatite activity and the mid-Zambezi-Luangwa rift zone in Northern Rhodesia that is the concern of the present paper.

## II. The Mid-Zambezi-Luangwa Rift Zone

The main elements of the rift in Northern Rhodesia are shown in simplified form in Text-fig. 2, in which it can be seen to extend right across the Territory from the Tanganyika border in the north-east to the Deka-Wankie area in Southern Rhodesia, a total distance of 800 miles. This plotting of the rift pattern was first prompted by the fact that the seven carbonatites so-far discovered in Northern Rhodesia are all localized along this zone of the crust : it is based on the mapping of the main Concession Companies in Northern Rhodesia (Guernsey, 1941 and 1950), more recent work by Gair (1959) and Tavener-Smith (1960) in the Zambezi valley, and the writer's own observations in many parts of the rift zone.

While attempting to elucidate the structural control of four of these carbonatites in the lower Luangwa-Rufunsa area (Bailey, 1960), it was noted that the rift lost much of its identity in the region between the north-eastern (Chirundu) end of the mid-Zambezi trough and the south-western end of mid-Luangwa trough. The stretch of country between these two colinear troughs is mainly an upland region of Pre-Cambrian metamorphics and granite, bounded on the north by the E.-W. Luano rift valley, and on the south by the E.N.E.-W.S.W. portion of the Zambezi trough between Chirundu and Mpata Gorge ; the Luano trough representing a westerly extension of the mid-Luangwa rift and the Chirundu-Mpata trough an easterly extension of the mid-Zambezi rift. These deflections in the N.E. trending rift zone are the most noticeable consequences of a powerful element of E.-W.




TEXT-FIG. 2.—The mid-Zambezi-Luangwa rift. Eruptive centres : (1) Rungwe ; (2) Panda Hill ; (3) Ilomba ; (4) Nkumbwa Hill ; (5) Mivula ; (6) Rufunsa ; (7) Mkwisi ; (8) Keshya.

fracturing intersecting the mid-Zambezi-Luangwa line. Most of the changes in trend and other complexities in this area can be explained in terms of interference of N.E.-S.W. and E.-W. fracture patterns, with one or other tending to predominate in different parts. Inspection of a wider region reveals that the area is, in fact, at the intersection of the mid-Zambezi (N.E.-S.W.) and lower-Zambezi (E.-W.) rifts (see Text-fig. 2). Details of the pattern of the lower-Zambezi rift in Portuguese East

Africa are not known to the writer and it has been merely sketched in the diagram (Text-fig. 2) but it is interesting that it too " loses its identity " at the intersection with the mid-Zambezi rift, reappearing, almost certainly, in the Kafue Flats basin to the west: Guernsey (1941) in fact suggested that the basin is part of the Northern Rhodesian rift system " deflected away from the normal north-east-south-west direction ".

## III The Associated Carbonatites

One carbonatite, Nkumbwa Hill (Reeve and Deans, 1954), is in the upper Luangwa valley (Text-fig. 2) and its relation to the rift is apparent, lying close to the north-west marginal zone : it is probably much greater in age (? Pre-Cambrian) than the other Northern Rhodesian carbonatites (personal communication T. Deans).

The other six carbonatites are located in the complex area around the intersection of the mid-Zambezi and lower-Zambezi rifts. Four of these, in the Rufunsa valley (Bailey, 1960), form the outstanding features of an unusual carbonatite volcanic province at the intersection of the south-east marginal fracture-zone of the Luangwa rift and the small Rufunsa rift, trending east-south-east. At the close of Karroo times (early Jurassic) the major Karroo boundary faults were initiated during an important phase of rift-movement (Dixey, 1955, p. 39) in which the present troughs were formed. In the course of these movements, in the Rufunsa area, (Text-fig. 3) there was intermittent local buckling of the Karroo rocks at the rift intersection and a large concordant, composite intrusion of carbonatite (Kaluwe) was emplaced in a syncline in the Upper Karroo sandstones. Renewed movements during early Cretaceous times triggered off a new phase of activity with the formation of several volcanoes erupting carbonatite ash and agglomerate, and into the three largest vents plugs of carbonatite marble were intruded. It seems clear that this rift intersection formed a critical zone of crustal weakness which was exploited by carbonatite fluids making their way to the surface from great depth.

About 100 miles west of the Rufunsa province two other carbonatites are found in the valleys of the Keshya and Mkwisi streams (Bailey, in press). The Keshya carbonatite is an irregular intrusion of coarse calcite-marble and breccia into the roof-migmatite zone of the Kafue Gorge granite of Katanga age (Late Pre-Cambrian–Early Palaeozoic). Twenty-one miles to the north the smaller, but essentially similar Mkwisi carbonatite intrudes Katanga metasediments. As they are found in an upland area of Pre-Cambrian rocks the relationship of these two carbonatites to the rift pattern is not as immediately apparent as was the case at Rufunsa : consideration of Text-fig. 2, however,

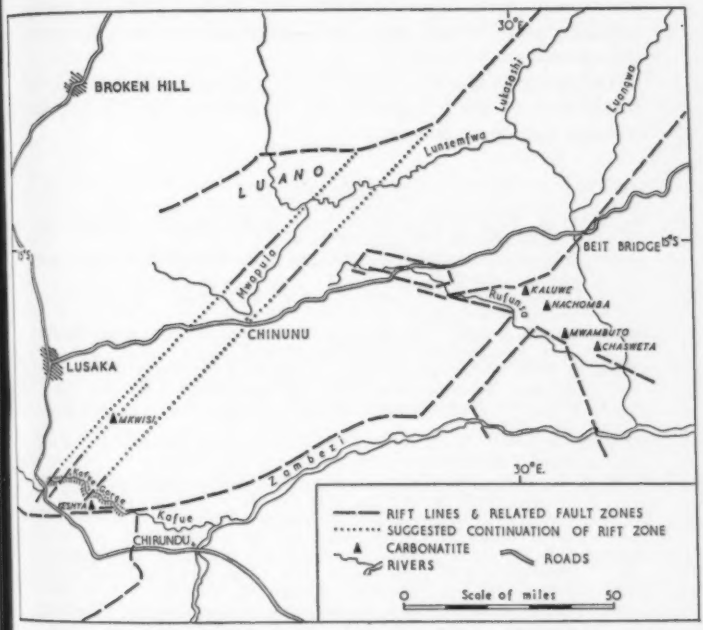
shows clearly that they are in the zone of intersection of the mid-Zambezi and lower-Zambezi rifts, and with the aid of Text-fig. 3 it is possible to examine the relationship more closely.

Consider first the Mkwisi carbonatite. In the detailed mapping of this intrusion no large fault or linear zone of dislocation came to light,[1] but this is not altogether surprising because the area covered was small and there is no detailed information on the geology of the surrounding country : the regional position of the intrusion is highly significant, nonetheless, for it lies on the line of the north-west marginal fracture-zone of the Luangwa rift. This rift at its south-west end is, in effect, a double rift, the parallel Luangwa and Lukasashi troughs, and it is the north-west escarpment of the latter which swings west to form the northern escarpment of the Luano valley (Text-fig. 2). But the N.E.–S.W. lineament of the Lukasashi does not die out with this westward deflection, it is continued across the Luano valley into the Mwapula valley (Text-fig. 3). This latter valley is a small rift-trough and the south-westerly extension of the disturbance is almost certainly marked by the strong hot-spring activity at Chinunu, on the Great East Road. Between Chinunu and the Kafue Gorge, fifty miles to the south-west, lies a little known area of Pre-Cambrian metasediments and granite, but it is almost certain that the south-west extension of the Lukasashi-Mwapula zone of disturbance is marked by the strong, colinear N.E.–S.W. faults of the upper part of the Kafue Gorge (Text-fig. 3, see also Bailey, 1955). These faults control the step-like plan of the Kafue river's course and in fact cause its sudden change of direction from south-east in the lower Kafue Flats to north-east where it plunges into the Gorge. It is surely not coincidence that the Mkwisi carbonatite, between the Mwapula rift and the N.E.–S.W. fault zones of the upper Kafue Gorge, lies exactly on the lineament of which they are parts, namely the south-west extension of the Luangwa-Lukasashi rift-zone.

In the case of the larger Keshya intrusion the relationship to the rift pattern is easily explained : it lies on, but largely to the north of a big fault-zone, trending north of east, which is followed by the Keshya ravine. This fault is part of an E.-W. echelon of faults affecting the lower Kafue Gorge, and dominating the fracture pattern of the region to the east to such an extent that the escarpment of the mid-Zambezi rift swings east-north-east when it reaches the Kafue, in response to this trend (Text-figs. 2 and 3). This E.-W. component of faulting is simply the westward continuation of the lower-Zambezi rift-zone, and it is carried through by the faults of the lower Kafue Gorge, the Keshya fault, and still further west, the southern boundary fault of the Karroo

[1] A possible indication may be the existence of a warm spring just to the south.

sandstones and lavas at the eastern end of the Kafue Flats. As described by Guernsey and shown on Text-fig. 2 this last zone of dislocation is traceable at least as far west as Lochinvar ranch on the south side of the Flats and recent seismic disturbances (Guernsey, 1941, p. 106–107) indicate continued movement to the present day.




TEXT-FIG. 3.—Simplified plan of the rift pattern between the mid-Zambezi and Luangwa troughs.

## IV Conclusion

It has been shown that the six carbonatites in the southern part of Northern Rhodesia can all be related to the complex tectonic disturbances around the intersection of the mid-Zambezi and lower-Zambezi rifts. A similar relationship holds at the intersection of the Luangwa rift and the Nyasa rift where the Rungwe volcanic centre is located, as pointed out by Dixey (1955, p. 28). In addition the Panda Hill carbonatite and related complexes, and possibly too the Nkumbwa Hill carbonatite, and the Ilomba and Mivula alkaline syenites (Bloomfield, 1959 ; Newton, 1959) represent older phases of activity around this intersection (Text-fig. 2).

It seems significant that two important foci of igneous activity, in particular alkaline and carbonatite activity, on the Gobabis-Ruaha rift-zone (Text-fig. 1) are found at its intersections with other rifts. Whatever may be the cause of rifting, therefore, such intersections must be particularly favourable for perforation of the continental crust by deepseated igneous fluids : probably the intensive fracturing creates vertical zones of tension, which have a more effective penetration into the sub-crust at these points.

The author is indebted to Mrs. K. M. Davies for assistance in the preparation of the text-figures and to Professor W. D. Gill for critically reading the transcript.

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# The Age and Relationships of the Cornbrook Sandstone

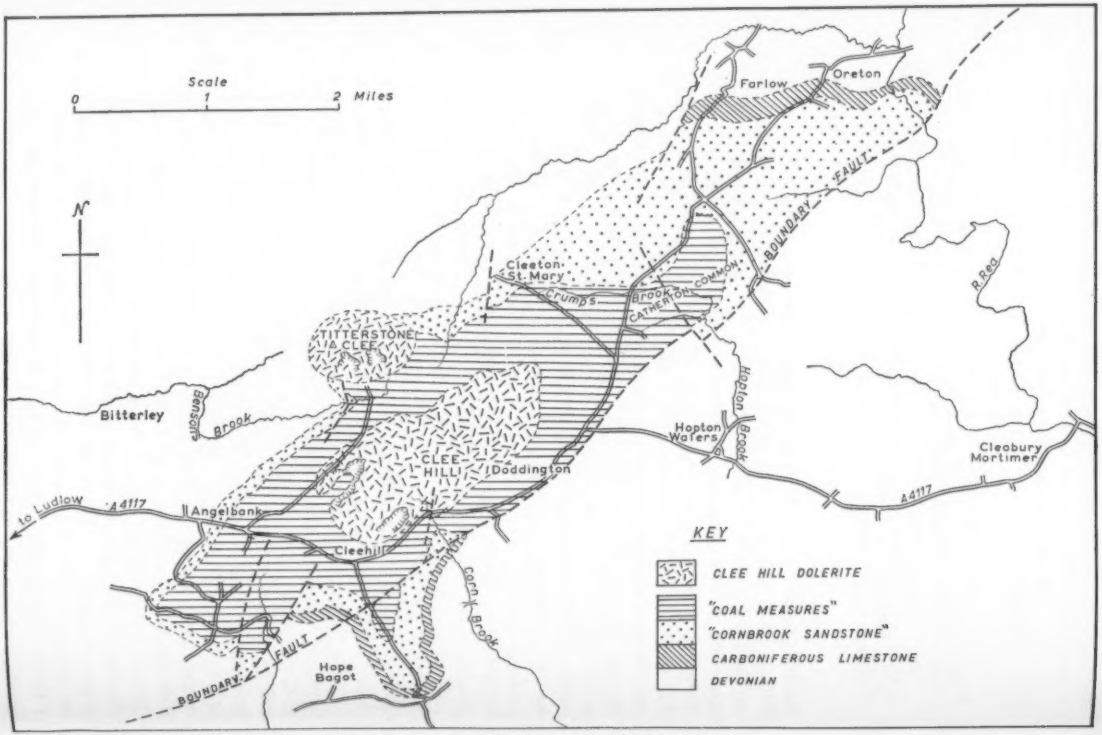
By D. G. Jones and T. R. Owen

Abstract

The Cornbrook Sandstone of Titterstone Clee Hill has plants of Westphalian (Middle Coal Measures) age. It is thus probably a correlative of the lower Kinlet Group of the Forest of Wyre. The base is unconformable to the Old Red Sandstone and Carboniferous Limestone while the top is apparently conformable with Productive Coal Measures of the Sweet Coal Group, also of Middle Coal Measures age.

## Introduction

The formational name " Cornbrook Sandstone " was given by Dixon (1917, p. 1064) to a series of grits and conglomerates with subordinate clays and thin coals on Titterstone Clee Hill in Shropshire, lying between Old Red Sandstone and Carboniferous Limestone below and Productive Coal Measures above. This followed from a suggestion of Vaughan (1905, p. 254), that the beds of 1,000 feet maximum thickness, were analogous in stratigraphical position and age to the Avonian Drybrook Sandstone of the Forest of Dean and Bristol areas. Previous workers such as Prestwich (1840, p. 413), and Jones (1871, p. 363), had used the term " Millstone Grit " for the formation. Dixon believed that it rested conformably on a single clay horizon of the Carboniferous Limestone and, by simple overlap, unconformably on the Old Red Sandstone. An Avonian age for the formation was also attested by the description of Kidston (1917, p. 1076) of apparently Lower Carboniferous plants at a few localities. The lowest beds of the overlying Coal Measures were further regarded by Kidston as of Lanarkian age and were believed by Dixon (1917, p. 1067) to rest with angular unconformity on the Cornbrook Sandstone. George, (1956, p. 308) on the basis of discrepancy in dip and strike suggested that the Cornbrook Sandstone unconformably transgressed various horizons of the Carboniferous Limestone. These limestones, he showed, were correlatives of the Lower Limestone Shales and Oolite Groups of Breconshire. The Cornbrook Sandstone, however, he allocated to the Namurian, it having apparently similar stratigraphical and structural relations to the limestones as the Millstone Grit of the North Crop in South Wales, 50 miles from the Clees. As the true Namurian is only 90 feet thick at the nearest point in South Wales, any thick Namurian in Shropshire would be in an anomalous position to the general outcrop. The authors, therefore, undertook this investigation of the age of the Cornbrook Sandstone. The study was extended to include some general structural and stratigraphical relationships of the formation.




TEXT-FIG. 1.—General geology of Clee Hill.

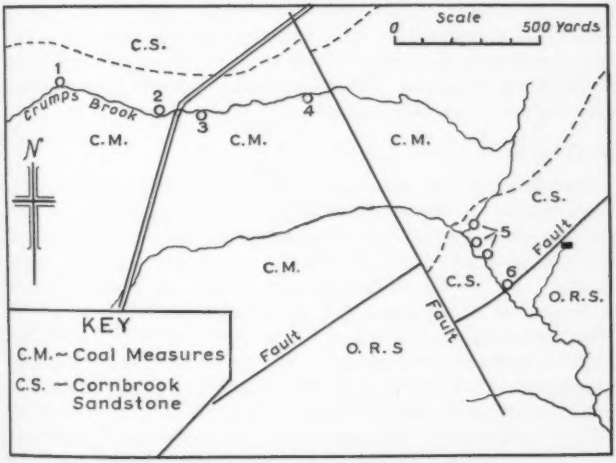
## Main Exposures

### *General*

The Carboniferous rocks of Titterstone Clee form part of a N.E.–S.W. trending syncline, probably fault-bounded on its south-eastern side. Although extensive exploitation of the sequence has proceeded at many levels for limestone, ironstone, coal, and building stone, exposures are few and poor. There are in fact only three reasonable sections—(*a*) Hopton Brook, (*b*) Cornbrook Dingle, and (*c*) Benson's Brook (Text-fig. 1).

### *Details* (*a*) *Hopton Brook*

The outcrops in this section are shown in Text-fig. 2. Between localities 1–2, the dip is mainly S.S.E. to S.E. at 5° to 15°, with frequent




Text-fig. 2.—Upper portion of Hopton Brook.
(For explanation of locality numbers, see text.)

sharp rolls. Incompetent folding is present and is more marked in the ferruginous underclays exposed at intervals in the stream. The main portion of the sequence is composed of light brown to grey or yellow sandstones, often with " espley " type bands, occasionally flaggy and with internal corrugations. Carbonized plant debris is frequent throughout. Some of the beds are similar lithologically to the Cornbrook Sandstone, but the presence of associated foetid sulphurous coals i.e., " stinking coals " and rashings (localities 3 to 4), places them in the Coal Measures proper lying above the Cornbrook Sandstone and

probably correlative with the Highley Group. Downstream near Knowlebank, at locality 5, sandstones and grits of Cornbrook Sandstone type lithology outcrop at intervals with a gradual southward increase in dip (mainly N.N.W.). They are massive yellow and brown sediments with appreciable beds of underclay and frequent coal smuts having an exposed thickness of near 400 feet. There is ample evidence of faulting, the beds sometimes steepening to vertical. At locality 6, the Cornbrook Sandstone is apparently faulted against Downtonian marls. The relationship of the lower limit of the Cornbrook Sandstone is thus indeterminate—see Dineley and Gossage (1959, p. 235)—but the outcrops near this eastern boundary fault are probably near the base as the sandstones become increasingly pebbly downstream. Although no palaeontological marker horizons are present in Hopton Brook, the sequence can be divided into three broad lithological units. A lower, mainly arenaceous, group—the " Cornbrook Sandstone "—with a thickness of at least 400 feet is succeeded by measures composed of thin coals, underclays, and flaggy sandstones. These are followed by an upper mainly arenaceous division with some thick underclays, coals, and conspicuous nodular ironstone.

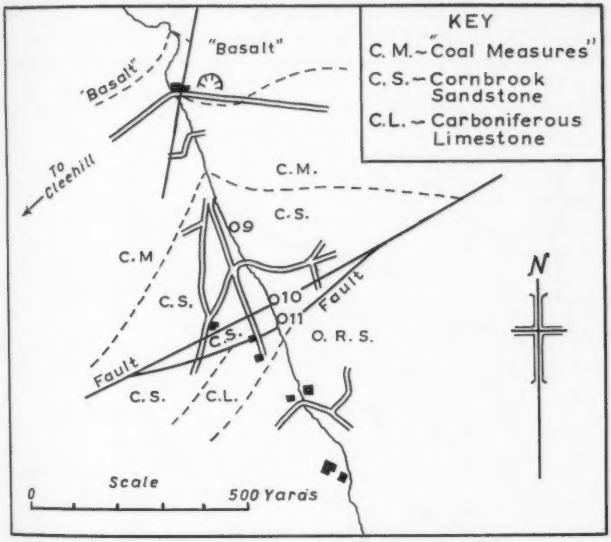
(*b*) *Cornbrook Dingle*

The succession here occurs in a steep gutter, shown in Text-fig. 3, where massive sandstones and thick clays are moderately well exposed. The sandstones are brown to orange coloured, frequently pebbly and with abundant carbonized plant-debris. Numerous gaps occur in the sequence and are due to the presence of clay bands which occupy hollows in the stream banks. Where exposed, the clays are grey to green, frequently mottled and often true underclays with rootlets and capped by coal smuts or rashings. The plant bed in Text-fig. 3, locality 9, lying about 250 feet below the local top of the Cornbrook Sandstone consists of thin grey mudstone underlying a thin coal, all sandwiched between massive grits. The upper limit of the Cornbrook Sandstone is not easily determined. In places it is overlain by coal measures, in others by the transgressive " basalt ". The lower beds of the sandstone, involved in the Eastern Boundary Fault zone, occur as fault slices at localities 10 and 11. In this, the type section, the true thickness of the formation is thus again indeterminate. The exposed thickness reaches 700 feet. Dixon (1917, p. 1067) gave a minimum of 1,000 feet, but this is considered excessive as repetition by faulting in the lower beds undoubtedly inflates this estimate.

(*c*) *Benson's Brook*

Exposures here begin some 60 feet below the " basaltic " sheet, horizontally overlying Productive Coal Measures of the Sweet Coal

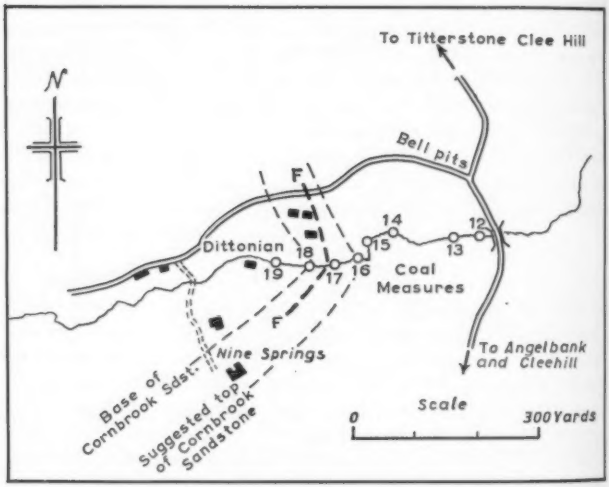
Group. The outcrop of a coal seam (probably the Great Coal of Dixon) is marked by old " bell " pit workings near the road in Text-fig. 4. Below this coal at locality 12, grey and green underclays with flaggy sandstones dip westward at 5° and contain well preseved plants—listed on pp. 292–3. Downstream to locality 16, the succession is made up of thick underclays, coals, and coaly rashings with sandstone bands




TEXT-FIG. 3.—Upper portion of Cornbrook Dingle.
(For explanation of locality numbers, see text.)

of " espley " type. Nodular ironstone is abundant in grey mudstone and also contains determinable plants at locality 15. The sandstones, some closely resembling the underlying Cornbrook Sandstone, show variable dips—both in amount and direction, due mainly to small-scale rolling. Toward locality 16 (at an iron fence) the dip gradually increases to 35° and is to the east. Farther downstream the sequence is continued, with some gaps, in sandstones with associated red or mottled marls and coal smuts. Dips are again variable, but average 20° to 25° to the east. At locality 17 the marls can be seen in the stream bed to overlie thick and current-bedded yellow grits with " espley " type bands all having a 20° to 25° eastward dip. A prominent bluff on the west of an old quarry (near loc. 17) marks the outcrop of massive yellow pebbly

grits (Text-fig. 5) dipping east at 45° to 50°. These are also well exposed in the stream and form a waterfall at the base of which (loc. 18), they overlie, disconformably, cream-coloured grits of the Dittonian. The basal beds of the Cornbrook Sandstone are massive conglomerates and have large pebbles, up to 10 inches in length, of vein quartz and orthoquartzite. Pebbles of cleaved slate, Old Red Sandstone and rarer Carboniferous Limestone also occur. The limestone pebbles are similar in lithology to the Oolite Group (Z age) of Titterstone Clee.

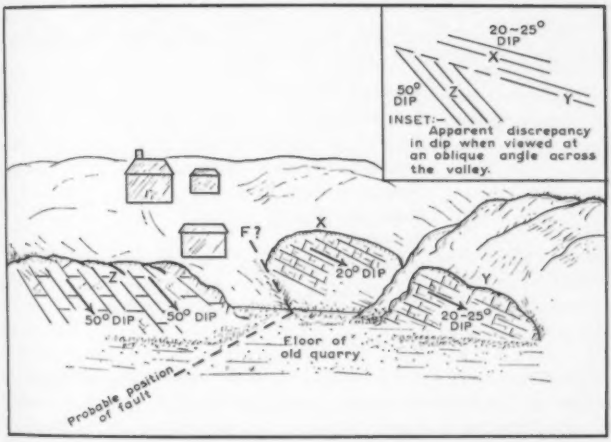



TEXT-FIG. 4.—Position of exposures in upper part of Benson's Brook. (See text for details.)

They contain cryptostomatous polyzoa and ribbed brachiopods such as *Camarotoechia mitcheldeanensis*. The conglomerate " trenches " the Dittonian grits, but the difference in strike is only slight and that of dip only 2° to 3°. The channelling and current-bedding suggest a depositional direction, for the Cornbrook Sandstone, from the W.N.W. Downstream the dip increases and becomes vertical in Dittonian Sandstones at locality 18. This strongly suggests the presence of a nearby strike fault.

While the lower limit of the Cornbrook Sandstone is undoubtedly unconformable to the Dittonian, it is not possible, on the evidence in the stream section, to support Dixon's (1917, p. 1067) contention that the Productive Coal Measures rest unconformably on the upper limit

of the Sandstone. Lithological separation between Cornbrook Sandstones and Coal Measures is difficult, the same rock types occurring above and below the junction which the authors would place near locality 16. The angular discordance described by Dixon at the critical locality can be ascribed to faulting (near loc. 17, Text-fig. 4). Here in the south bank of an old quarry (Text-fig. 5), an oblique view gives an apparent angular discordance due to the presence of this fault. The course of the fault is marked by a thin crushed belt at the base of a gulley on the north side of the brook and at the western end of a north bank disused quarry.




TEXT-FIG. 5.—Diagrammatic sketch of Cornbrook Sandstone exposures in Benson's Brook (between localities 17 and 18 in Text-fig. 4).

The thickness, seen in Benson's Brook, of Productive Coal Measures is 250 feet, and of massive Cornbrook Sandstone-type lithology, about 180 feet.

Although its outcrop can be traced by boulders at some points, apart from these stream sections, little else can be seen of the Cornbrook Sandstone. At Cleeton St. Mary considerations of dip suggest an unconformable relationship to the Old Red Sandstone. On the southern and south-western sides of Titterstone Clee the Cornbrook Sandstone probably gradually oversteps the Carbonifrous Limestone (Text-fig. 1). At the same time a marked thinning of Cornbrook Sandstone-type lithology must occur in a westerly direction across this southern part of Titterstone Clee.

## Age of Plant Horizons

(*a*) *Cornbrook Sandstone.*

Kidston (1917, p. 1076) recorded three plants which he designated as "Lower Carboniferous". One only of these was taken as definite evidence of age, i.e., *Lepidodendron veltheimi*, the others, *Stigmaria ficoides* and *Sphenopteris* sp., having no zonal value. *L. veltheimi* was later shown by Crookall (1939, p. 18) to be merely an aggregate form and thus of no value as an index of either a Lower or Upper Carboniferous horizon. The Cornbrook specimen of *L. veltheimi* is furthermore fragmentary and poor. A Lower Carboniferous age, even in the sense of Kidston's floral break, is thus not conclusively proven.

The plants occurring approximately 250 feet below the top of the formation in Cornbrook Dingle (loc. 9) are :—

*Asterophyllites* sp.
*Calamites* sp.
*Alethopteris lonchitica* Schl.
*Lonchopteris rugosa* Brongn.
*Neuropteris tenuifolia* (Schl.)
*Sphenopteris* sp.
*Pecopteris* sp.
*Dactylotheca plumosa* Artis.
*Diplotmema* sp.
*Samaropsis* sp.
*Carpolithes* sp.

The majority of these forms are typically Upper Carboniferous (as defined by the goniatite zonal scheme) and occur in the Westphalian. *N. tenuifolia* and *L. rugosa* are both characteristic of Westphalian B, which is approximately equivalent to the non-marine lamellibranch zones of *Modiolaris* and lower *Similis-Pulchra*, i.e., between the Amman (= Katharina) and Cefn Coed (= *Aegir*) marine bands. In South Wales where the floral sequence for the Upper Carboniferous is most complete, *L. rugosa* is restricted to the middle of *Modiolaris* Zone. No specimens typical of a Lower Coal Measures (= Westphalian A) or Namurian age are present and there is little doubt that most if not all of the Cornbrook Sandstone is of Middle Coal Measures age.

(*b*) *Coal Measures*

Plants from the Coal Measures of Benson's Brook (loc. 13, Text-fig. 4) are as follows :—

*Asterophyllites* sp.
*Sphenophyllum myriophyllum* Crépin.
*Sphenopteris* sp.

*Neuropteris* cf. *subplicata* Kidston (= *N. plicata* of Newell Arber, 1917).
*Mariopteris latifolia* (Brongniart)
*Carpolithes* sp.
From loc. 15 :—
*Neuropteris gigantea*
*Cyclopteris* sp.

These forms also indicate a Middle Coal Measures age for the Clee Hill Sweet Coal Group, as noted previously by Newell Arber (1913, p. 419) and Kidston (1917, p. 1071). The presence of *Mariopteris latifolia* and *N.* cf. *subplicata* suggest, however, a higher horizon than the flora of the Cornbrook Sandstone. They are more typical of Westphalian C, i.e., of the upper *Similis–Pulchra* and *Phillipsii* Zones. Such an interpretation of horizon is thus in accord with a general *Modiolaris* Zone age for the Cornbrook Sandstone and also with the record of *Anthracosia* (*Carbonicola*) *aquilina* by Trueman (in Marshall, 1942, p. 24) from the Coal Measures above the basalt on Titterstone Clee.

## General Stratigraphy

A Middle Coal Measures age for the Cornbrook Sandstone of Titterstone Clee shows that it probably correlates with the basal Kinlet Group of the adjacent coalfield of the Forest of Wyre. It also accounts for the absence of "Millstone Grit" on Brown Clee Hill and the presence of patches of the Kinlet Group resting directly on Old Red Sandstone near Cleobury Mortimer noted by Dineley and Gossage (1959, p. 239). It resolves the anomalous presence of thick "Namurian" strata in South Shropshire lying on the immediate southern flanks of St. George's land and suggests that there was probably no connexion during Namurian times between the North Crop of South Wales and Shropshire. In fact any connexion was probably severed by general uplift in early Viséan times. The Cornbrook Sandstone was deposited on an uneven surface produced as a result of post-Tournaisian tilting or folding. Its rapid changes of thickness and internal thinning northwards and westwards being in accordance with similar variable behaviour of the frequently arenaceous and conglomeratic lower Kinlet Group as noted by Whitehead and Pocock (1947, p. 37).

## Acknowledgments

The authors wish to express their gratitude to the Director of H.M. Geological Survey for permission to examine material in the Kidston Collection, and also to the Curator of the Sedgwick Museum for the

loan of specimens from the Newell Arber collection. To the technical staff of King's College is also due the authors' appreciation for assistance with the preparation of the diagrams.

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# A Llandovery Shoreline from the Southern Malverns

By H. G. READING and A. B. POOLE

(PLATES XV–XVI)

ABSTRACT

The paper describes an unconformable junction of Upper Llandovery upon Malvernian in the Gullet Quarry of the South Malvern Range and discusses the evidence for and the significance of other comparable junctions in the Malverns. It is considered that Taconic movements are a factor in Malvern tectonics and that the importance of a " western boundary fault " to the Malvernian has been overstressed in the past.

## INTRODUCTION

SINCE the end of 1959 excavations of the west face of the Gullet Quarry in the Southern Malverns have exposed a junction of the Pre-Cambrian and Silurian. This quarry lies on the north side of the pass which separates Swinyard Hill from Midsummer Hill and is half a mile from the northern end of the main Cambrian outcrop of the region. So far as the authors can ascertain this is the first time a Malvernian-Silurian junction has been exposed in the Southern Malverns.

## DESCRIPTION OF EXPOSURE

In the western part of the Gullet Quarry the Malvernian consists of a complex of pegmatitic gniess and schists. The surface of the Malvernian dips westwards at an angle of approximately 60°. It is weathered and gullied, and is overlain by a band of conglomerate (Plate XV). This conglomeratic band varies in thickness from 1 inch to 1½ feet and contains well-rounded fragments ranging from 4 mm. to 80 cm. across, all showing a zone of weathering at the surface.

The fragments can be divided into three broad groups. The first is of highly altered doleritic rocks, varying in grain size and degree of alteration ; several showed considerable calcitization. The second group is of acid rocks of granitic or granitic gneiss types. The third group consists of pegmatites. All the above groups are well represented by similar rock types in the Malvernian of the surrounding area.

The matrix of the conglomerate is variable, ranging from a grey clay to a grey-green sandstone. Between some of the boulders and occasionally coating them are corals and brachiopods, some in the position of growth. (Plate XVI). The fauna has been examined by Professor A. J. Boucot who has indentified " *Coelospira* " *hemisphaerica* (J. de C. Sowerby) and *Stricklandia lens ultima* (Williams). These brachiopods occur together in $C_4$ of the Upper Llandovery at Llandovery (Williams, 1951).

Above this conglomeratic band 50 feet of grey-green micaceous siltstone and fine-grained sandstone are exposed, with intercalated

bands 2 or 3 inches thick of shelly sandstone and glauconitic limestone. These beds dip to the west at approximately 45°.

### Other Malvernian-Silurian Junctions

Phillips (1848) described the finding by his sister of a Llandovery conglomerate on the western slopes of Worcestershire Beacon in the Northern Malverns. This " Miss Phillips's Conglomerate " contained rolled and angular pebbles of Malvernian embedded in clay and surrounded by abundant fossil shells. The junction with Malvernian was exposed and is shown in a sketch by Phillips (1848, p. 67) to be vertical. Phillips had no doubt that this was an unconformable contact of Silurian upon Malvernian.

The unconformable nature of the Silurian was apparently confirmed by the section exposed when the first tunnel was driven through the Malvern Hills at Colwall (Symonds and Lambert, 1861). Fossiliferous Llandovery limestone and shale were found in what they took to be a fissure towards the western edge of the Malvernian. They published a sketch of this and of the western junction of the Malvernian with the Silurian. Symonds and Lambert evidently think, though they do not state, that the latter junction is stratigraphical. On the other hand they say that there is " evidence of great pressure and crushing " and a faulted junction would appear more probable.

Further evidence of the faulted junction was supplied by Robertson (1926) after the second Colwall tunnel was excavated. He found the " fissure " of Symonds and Lambert but showed that it was in fact a fault zone. He also demonstrated clearly that the western margin of the Malvernian was a faulted junction with the Llandovery.

Robertson also mentioned an exposure of Llandovery with Malvernian pebbles " 220 yards south-east of St. James's Church, West Malvern ". The same outcrop is described by Bennett (1942) in a quarry under a sycamore tree. It is not the original " Miss Phillips's Conglomerate " which lay further south and must have been removed during the nineteenth century. It is still visible (SO764459). Conglomeratic pebble bands alternate with compact siltstone layers and occasional thin shelly limestones. The range of species does not differ significantly from that of the Gullet Quarry though " *Coelospira*" *hemisphaerica* and *Stricklandia lens ultima* are are extremely rare.

Many other localities of Llandovery against Malvernian were visible during the nineteenth century. They are mentioned by Symonds (1880) in his little book " Old Stones ". They include several in the neighbourhood of the Wyche, at Cowley Park, in the Malvern tunnel, at Wind's Point, at Midsummer Hill and at Hollybush where Llandovery is said to have lain upon Hollybush Sandstone. There is no confirmation of these exposures, and it is not easy to judge the nature of their junctions.

On the one hand Symonds refers to the Llandovery breccia rather than a conglomerate and he may, therefore, have mistaken a fault breccia. On the other hand he undoubtedly believed them all to be unconformable and from his references to boulders and fossils in hollows between the boulders one would be inclined to think that some at least must be stratigraphical junctions.

Symonds also writes of a Cambrian Conglomerate " at the great quarry at the Gullet Pass ". This has been seen by no one else, and must have lain well east of the present western face of the quarry. Only 200 yards from the present face there is still an exposure of Lower Cambrian quartzite and conglomerate apparently lying upon Malvernian, though the actual junction is obscure.

There are in addition patches of Llandovery upon the main Malvern range. Groom (1899, 1900) discusses them, shows them on his map, and concludes that the junctions are all faulted. Some lie high up on the Herefordshire Beacon, some are recorded as lying on the eastern slopes of this hill ; but there is no evidence as to whether their junctions are stratigraphical or faulted, or even whether the Llandovery is in place. It is perhaps significant that Llandovery only has been found and higher strata are not recorded.

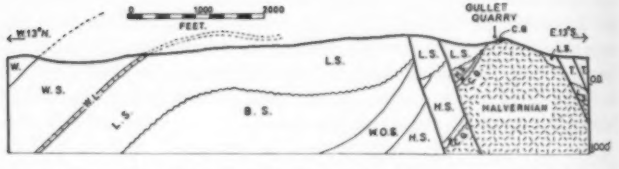
## Discussion

The most recent full description of the geology of the Malvern Hills is that of Groom (1899, 1900), who summarized his results in " Geology in the Field " in 1910. Groom maintained that the west boundary of the Malvern range is faulted and wherever the field evidence showed Silurian adjacent to Malvernian he mapped the junction as a fault. He thus implied that Cambrian is everywhere present below the Silurian and he denied that there was any evidence for transgression of Silurian upon the Malvernian. This idea of a major western boundary fault was supported by the Colwall tunnel section (Robertson, 1926).

Of recent years discussion of the tectonics of the Malverns has been reopened. Falcon (1947) queried both the age of the main folding movements and the presence of the " great Malvern fault " on the east side of the range. He claimed that the eastern margin of the Malverns was a pre-Triassic unconformity, modified locally by post-Triassic faulting. Among his evidence he used a comparison of the two junctions of the Malvernian seen in the tunnel section by Robertson. The Malvernian, against Silurian on the west side, is considerably less disturbed than the Malvernian, against the Trias, on the east. Falcon suggests that the greater disturbance of the Malvernian on the east is the result of pre-Triassic weatherng. Surely the simpler explanation of this evidence is that the western fault is one of lesser importance.

Further contributions to Malvernian tectonics have been made by Raw (1952) and Blyth (1952). Raw's thesis is that the Malverns are an Armorican nappe subsequently divided into strips by normal faults stepping down to the east. Blyth's view is of fault blocks thrust upwards in post-Lower Old Red Sandstone times. Both authors published sections across the Southern Malverns with a complete and uneroded Cambrian sequence and have shown the western boundary fault as a major fault, Raw postulating a throw of 4,000 feet.

Bennett (1942) also believed in the presence of a " western boundary fault ", but both he and Pocock and Whitehead (1948) state that the Silurian lies uncomformably upon the Cambrian and Pre-Cambrian. The latter base their opinion upon the occurrence of " Miss Phillips's Conglomerate ", and it is a view that follows the accepted nineteenth




TEXT-FIG. 1.—A section through the north side of the Gullet Quarry. T = Trias ; W = Wenlock Limestone ; W.S. = Wenlock Shale ; W.L. = Woolhope Limestone ; L.S. = Llandovery Sandstones ; B.S. = Bronsil Shales ; W.O.S. = White-leaved Oak Shales ; H.S. = Hollybush Sandstone ; C.Q. = Cambrian quartzites and conglomerate. Vertical scale = horizontal scale.

century picture. It is clearly put by Holl (1865) who goes further than Phillips and emphatically states that the Llandovery oversteps the various divisions of the Cambrian on to Pre-Cambrian and also that the higher division of the Llandovery overlaps the lower division.

The Gullet Quarry confirms this. In less than a mile the full Cambrian sequence of 2,500 feet of sediments and Ordovician intrusives is lost. The unconformity increases towards the Malvern range and so, on the assumption that the Cambrian at one time extended over the range, the Malverns must have risen in late Ordovician—early Silurian times, tilting and faulting the Cambrian to the west and allowing it to be removed.

A structural section is shown through the Gullet Quarry (Text-fig. 1). The writers envisage upward movement of the Malverns in both Taconian and Hercynian times, causing both folding and faulting of the Palaeozoic. Steep reverse faults were produced by this upthrusting, the faults turning into low angle thrusts, possibly of the nature of slip

sheets, on the flanks of the Malverns, e.g., below Herefordshire Beacon. Subsequently there was a period of normal faulting, stepping down to the east, as advocated by Raw (1952). Some of the later faults may have followed earlier fractures and so three periods of movement can be discerned along them.

## Conclusions

The view is vindicated that there is an unconformity of considerable magnitude below the Silurian, and that the Malverns were uplifted vertically in pre-Upper Llandovery times. This unconformity, as in South Shropshire, is probably Taconic. It must be allowed for in any structural interpretation of the Malverns.

Detailed structural analysis is not the purpose of this paper, but doubts have been cast upon the presence of a "western boundary fault", separating Malvernian from Lower Palaeozoic. Faulting undoubtedly exists; it has been proved in the Colwall tunnel and elsewhere, but it is not a continuous major fault like the "eastern boundary fault".

## Acknowledgments

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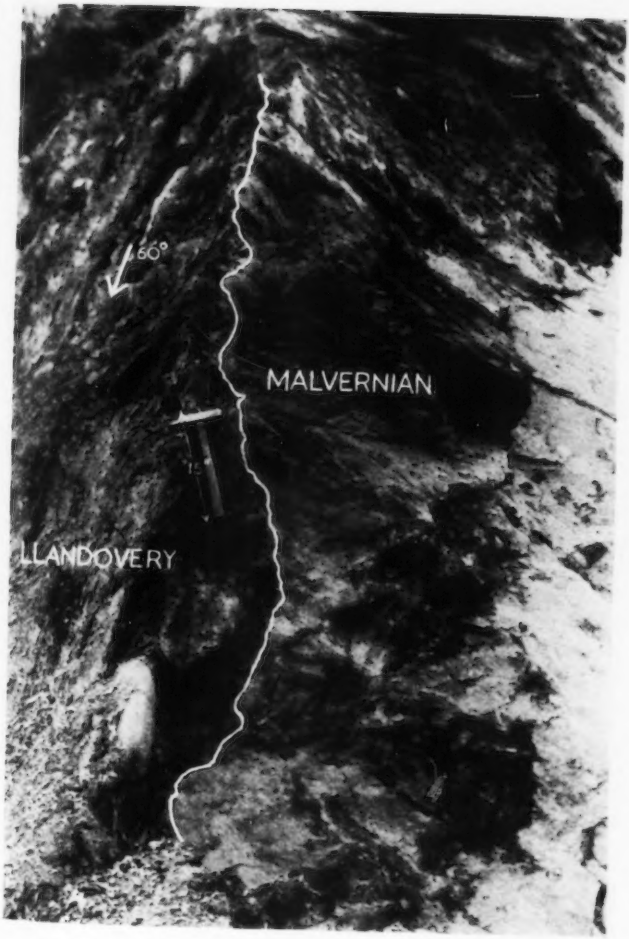
## EXPLANATION OF PLATES

PLATE XV.—View of the western face of the Gullet Quarry looking north.

PLATE XVI.—Part of a boulder from the conglomerate showing highly fossiliferous Llandovery siltstone adhering to its surface.




THE GULLET QUARRY.

*facing p.* 300.

LLANDOVERY CONGLOMERATE FROM THE GULLET QUARRY.

# Rock Magnetism Applied to Some Geological Problems

By D. J. BLUNDELL

ABSTRACT

Recent studies concerned with the application of rock magnetism to several geological problems serve to illustrate the scope of the methods at present available. Rock magnetism can usefully be applied to determine the age of a rock or to correlate it with some other. It may be used in studying the structural history of a formation or the thermal history of a rock. The examples cited deal with igneous rocks in Britain, but the methods also apply to some sediments, and possibly to metamorphic rocks.

It is stressed that rock magnetism can be applied quite independently of any theory concerning the earth's field. It is necessary only to establish whether the magnetism measured in the laboratory is a true record of that acquired when the rock formed, or at some other known time in its history.

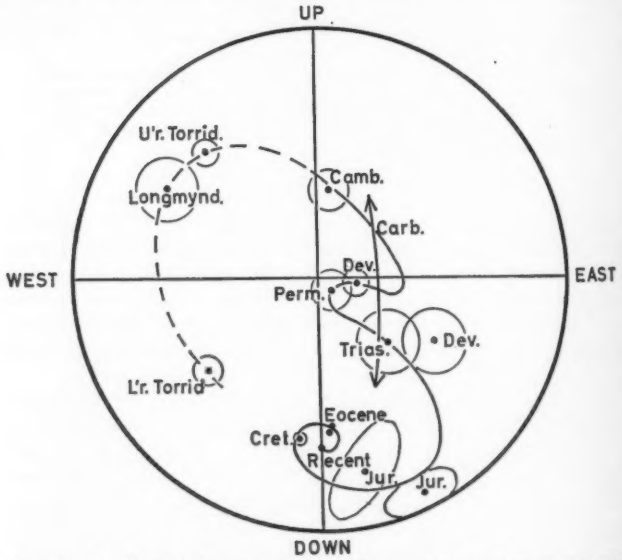
## INTRODUCTION

IN RECENT years attention has been paid to the study of the magnetic properties of rocks as a means of aiding the interpretation of geological events. Such problems as the age, attitude, or history of various formations may be resolved and it is the purpose of this paper to outline the methods available at present and to give specific examples of their application. The examples given are concerned with formations in Britain although similar studies are being carried out elsewhere.

It is observed experimentally that if a rock cools from above the curie point of its ferromagnetic constituent (generally 500°–600° C.) in a weak magnetic field such as the earth's, it acquires a permanent magnetism in the direction of the field. This thermoremanent magnetism (T.R.M.) is generally very stable. An igneous rock, cooling during formation, can thus acquire a permanent magnetism in the direction of the prevailing earth's field, which is retained for millions of years. A sediment can become magnetized during deposition as the ferromagnetic particles tend to become aligned in the direction of the prevailing field. There are often stronger forces acting on the particles during sedimentation which disturb the alignment, however, so the direction of magnetism of such a sediment does not always reflect exactly the prevailing field direction. A rock can also become magnetized during a chemical change, as, for example, the transformation of goethite to hematite, which may take place either during or after formation. This magnetism has been shown to be stable and directed along the prevailing field, and it is known as chemical remanent magnetism (C.R.M.). Many red sediments are magnetized in this way.

Only if it can be shown that the remanent magnetism of a rock, as measured in the laboratory, preserves a true record of the direction of the field prevailing during formation or at some other known point in

the rock's subsequent history, is it possible to make proper use of the information. Fortunately there are a number of tests of magnetic stability available by which rock samples with stable magnetism can be separated from those with unstable magnetism. A good description of these tests has recently been given by Cox and Doell (1960, pp. 655–661). Quite often a rock has unstable components of magnetism superimposed on a stable component, but it is possible to remove the




TEXT-FIG. 1.—Simplified stereographic representation of past magnetic field directions in Britain (data from Cox and Doell, 1960). North hemisphere stereogram.

unstable parts by partially demagnetizing the rock in an alternating magnetic field. This method of " cleaning " samples widens the range of rocks which can be used but the difficulty in finding samples with stable magnetism can sometimes be a disadvantage.

Measurements have been carried out over the past ten years to determine the directions of magnetism of a wide variety of rocks of known age in the British Isles, throughout the geological succession. A most comprehensive list of these is given by Cox and Doell (1960). From these it has been possible to build up a picture of the direction of the magnetic field in Britain prevailing at various times in the past. Text-fig. 1 shows in a stereographic projection how there has been a

slow change in the field direction from practically horizontal during the lower Palaeozoic to the present field dipping downwards at about 68°. There is also a short period change, the secular variation, with a period of 500 years or so which is not apparent in the diagram since the points represent data in which this effect has been averaged out. Secular variation can account for departures of up to 15° from the present field direction. A wide sampling of a formation is generally adopted to average out the effects of local magnetic anomalies and secular variation. In some cases, however, the effect of secular variation can be used to advantage. A statistical method of analysis introduced by Fisher (1953) is very useful in gauging the scatter of directions of magnetism of individual samples from a single formation or locality and allows tests of significance to be made.

A feature of the magnetism of many rocks is that the polarity appears to be reversed. This has been attributed to periodic reversals of the polarity of the earth's main field but in a few cases it has been demonstrated that it is due to some mineralogical property of the rock, which produces a remanent magnetism in the direction opposed to the applied field. It is generally considered that both processes occur, though it is often not necessary to know the cause of reverse polarity to be able to make use of the phenomenon, and lack of knowledge of the polarity of the past earth's field is usually no disadvantage.

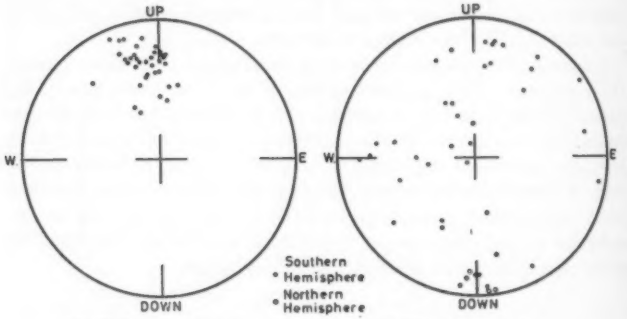
## Examples of the Use of Rock Magnetism

In order to give some idea of the ways in which rock magnetism can be of help to the geologist several examples are now briefly described where rock magnetism studies have produced an increased knowledge of local geological events. They are grouped under four headings.

(*a*) *The determination of the age of an intrusive suite.* The mean direction of the remanent magnetism of a suite of rocks of unknown age can be compared with the directions of magnetism of various rocks of known age, for instance those shown in Text-fig. 1. In this way it has been possible to infer the age of a dyke swarm on Lundy (Blundell, 1957). Lundy is made up almost entirely of a composite granite boss, and intimately associated with this is a dyke swarm. The granite closely resembles the Permo-Carboniferous granites of South-West England but the dyke swarm has petrological affinities with the Tertiary igneous complexes of North-West Scotland. The directions of remanent magnetism of forty-one samples taken from fourteen of the dykes were measured. As a result of storage stability tests two groups of directions were recognized (Text-fig. 2), one with stable, the other with unstable magnetism. The magnetism of the stable group has a mean direction 17° W. of S. declination and 59° upward inclination. Clearly the

magnetism is reversely polarized but allowing for this and comparing its reverse direction with those shown in Text-fig. 1 it is evidently closest to the direction of magnetism of Tertiary Rocks (the Antrim Plateau Basalts in particular) and is more than 30° from the Permo-Carboniferous direction. Using Fisher's statistics it can be shown that this is a significant difference. Here is circumstantial evidence that the Lundy dyke swarm is Tertiary in age rather than Permo-Carboniferous.

A similar situation occurs in Ayrshire. Among the minor intrusions there are two groups, the " kylites " and the " crinanites " which were at one time thought to be differentiates of the same magma, but later



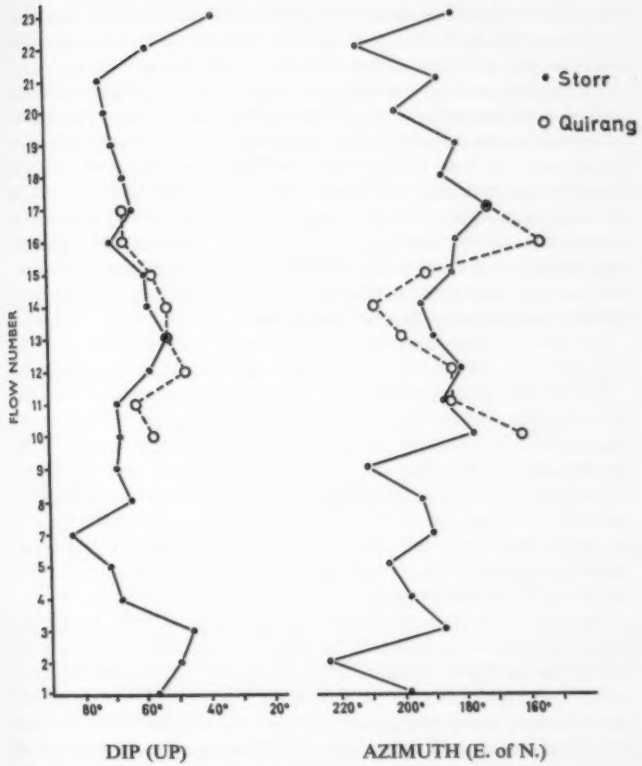

(A) Stable specimens. (B) Unstable specimens.

TEXT-FIG. 2.—Lundy dyke swarm (Blundell, 1957). Equal area projection.

work suggested otherwise. It is fairly well established that the kylites are Permian, but the field evidence regarding the crinanites is less conclusive. More than forty samples were collected from the two groups by Armstrong (1957) who found the kylites to be magnetized due south with 7° downward inclination, thus verifying their Permian age, while the crinanites were magnetized in a direction more consistent with their being Tertiary in age.

It must be pointed out that the ages of the Lundy dyke swarm and the Ayrshire intrusions could be worked out using rock magnetism only because of the large age differences involved. It is unlikely that the method can ever be used with success in stratigraphy as its resolution is so poor, but in principle there is no reason why it should not be equally applicable to sedimentary rocks.

(*b*) *The correlation of lava flows in different localities.* The possibility of making use of the effect of secular variation has already been mentioned. Khan (1960) has applied this effect in correlating piles of Tertiary lava flows outcropping in two exposures ten miles apart on the




TEXT-FIG. 3.—Magnetic directions of Tertiary lava flows from Storr and Quirang, Skye. (Khan, 1960. Reproduced by permission of the author).

Isle of Skye. From the field evidence it was clear that the flows in the two exposures belonged to the same group, but exact correlation could not be achieved. Khan measured the direction of magnetism of every flow at the two exposures and plotted magnetic azimuth and inclination against flow number (Text-fig. 3). At Storr, where twenty-two flows were sampled, a definite pattern emerged, due presumably to the effect of secular variation. Eight flows were sampled from Quirang, and Khan " matched " the pattern obtained there against that at Storr, working out correlation coefficients obtained by equating the basal Quirang flow with each of the lower fifteen Storr flows in turn. He

obtained a significant result only when the basal Quirang flow was equated to the tenth flow at Storr, and thus concluded that the eight flows at Quirang correspond to flows ten to seventeen at Storr.

Applications of this method are likely to prove widely useful when mapping lava formations. Use has already been made of the occurrence of sequences of reversed and normally magnetized lava flows in Iceland (Einarsson, 1957) as a means of correlating flows from one area to another, the changeover from normal to reverse magnetization acting as a kind of " marker horizon ", and it may well find application in this country. A knowledge of the intensity and direction of the remanent magnetism is also becoming valuable in the interpretation of aeromagnetic surveys carried out over igneous rock formations, particularly where reversed magnetizations are found.

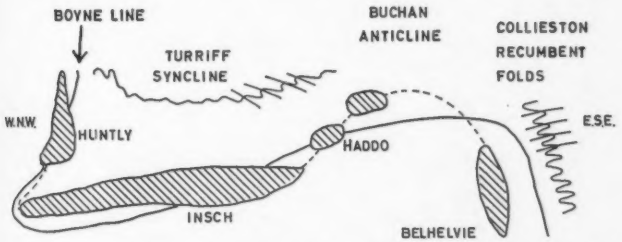
(*c*) *The determination of the structural arrangement of the Aberdeenshire gabbros.* A stable remanent magnetism remains in the same direction relative to the rock, so when a formation becomes tilted the magnetic direction is tilted accordingly. Thus directions of magnetism of sedimentary rocks are always measured relative to the bedding planes. The idea can be used to determine the extent of post-formational tilting of igneous bodies, where there may be ambiguous geological evidence as to their original attitudes. For instance, it is sometimes possible to tell whether the dip of a lava flow is original or post-formational. Khan (1960) has been able to infer in this way that the lavas of Skye were extruded over surfaces sloping at up to 15°.

An interesting example of the structural use of rock magnetism concerns the possible deformation of the Younger Basic intrusions of North-East Scotland. These form six large and many smaller bodies of similar character, and there is little doubt that they all belong to one igneous event, and possibly form a continuous sill-like body. In three of the main masses, Huntly, Insch, and Belhelvie, there is evidence of banding. It is near vertical at Huntly and Belhelvie and near horizontal at Insch. Shackleton (discussion of Read and Farquhar, 1956) has considered the banding to be due to gravity differentiation of the magma during formation and to be originally nearly horizontal, and he put forward the idea of a large scale post-gabbro folding to upend the Huntly and Belhelvie masses, together with, of necessity, the formation of the Buchan anticline (Text-fig. 4). Read (Read and Farquhar, 1956) was unable to accept the post-consolidation folding of the gabbros, which abound in delicate textures that would have been destroyed by such movement, and maintained that the Buchan anticline formed before the intrusion of the gabbros.

The application of rock magnetism (Blundell and Read, 1958) seemed a way of resolving the controversy, for had the masses tilted relative to each other the magnetic directions would have done likewise.

Consequently each of the major masses was sampled and magnetic directions were measured. Stringent tests for magnetic stability were applied together with microscopic analysis of thin sections. Of the sixty samples collected, twenty-one satisfied the requirements of the tests and were used in the final analysis. For each sample used it was demonstrated that the magnetism was stable and that the rock was chemically and mechanically unaltered.

The result (Text-fig. 5) was that the rocks from every mass had nearly the same direction of magnetism. Had the tilting envisaged by Shackleton occurred the magnetic directions of the Huntly and Belhelvie masses would have been 70° away from that for the Insch mass. Statistically there is no significant difference between the magnetic directions for the various masses, so from the magnetic



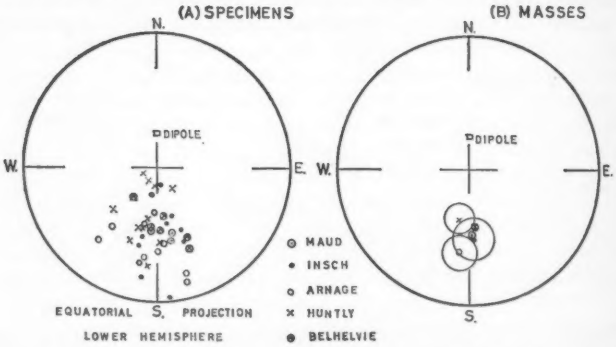

TEXT-FIG. 4.—Professor Shackleton's interpretation of the attitude of the Basic Masses of North-East Scotland (reproduced by permission of the Council of the Geological Society of London from the discussion to Read and Farquhar, 1956).

evidence it would appear that the folding suggested by Shackleton is unlikely to have occurred, and that the masses have remained relatively undisturbed since their formation. The mean direction of magnetism of all the masses is in broad agreement with those of Lower Palaeozoic rocks elsewhere in Britain and congruent with the intrusion of the Aberdeenshire gabbros as part of the Caledonian orogeny.

(*d*) *The magnetic properties of country rocks locally heated by an intrusion.* In effect, each of the previous examples has made use of the remanent magnetism as a fossil, to correlate one rock with another. Further geological applications of rock magnetism make use of the nature of the remanent magnetism as well as its property of great stability. For example, when a rock is heated, say by a nearby later intrusion, it acquires a new magnetism as it cools which is characteristic of the magnetic field then prevailing. If the heating is to a temperature above the curie point then all the original magnetism is destroyed and

a new T.R.M. is produced. If, however, the heating is to some intermediate temperature then not all the original magnetism is lost and the rock acquires instead a partial thermoremanent magnetism (P.T.R.M.). It is a property of P.T.R.M. that it can be completely destroyed by reheating to the maximum temperature from which it was acquired, leaving only any of the original T.R.M. not destroyed in the heating.

Everitt (1960) has made use of the first of these properties in an investigation of the Midland basalts. These form a series of some half-dozen sheet-like bodies and doubt remained as to whether they are sills or lavas. Pocock (1931) maintained that exposures at all the sites are of lava, with the exception of the " Pouk Hill " body. Later work

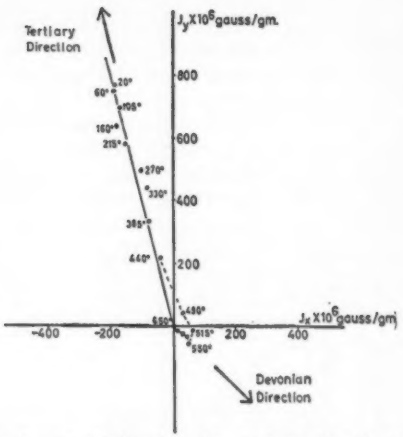



TEXT-FIG. 5.—Equal area projections showing magnetic directions for (A) stable specimens (B) mean directions for the Masses, with circles of 67 per cent confidence. (Blundell and Read, 1958. Reproduced by permission of the Council of the Geologists' Association).

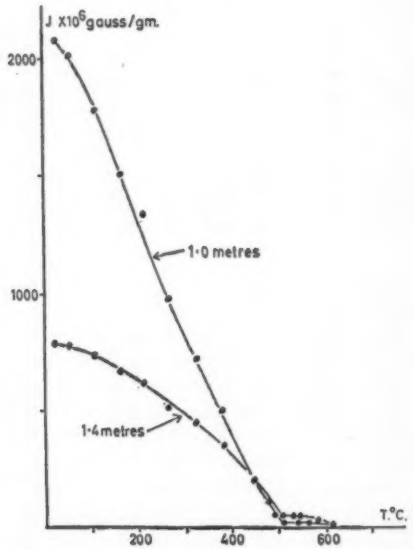
led Marshall (1942) to agree on the extrusive nature of the rock at Little Wenlock but to consider all the other bodies as intrusive, though the evidence was dubious at Kinlet. Everitt collected samples from the exposures in question both of the igneous rock and the country rock, and measured their directions of magnetism summarized in Table 1. He found that at each exposure except Little Wenlock the country rock in contact with both the bottom and the top of the igneous body had a magnetism characteristically of thermoremanent origin, much stronger and less scattered than that of the country rock some distance from the body. He concluded that the country rock had been locally heated by the igneous body, having thus acquired a T.R.M., and since both bottom and top contacts were similarly affected the body must be a sill. At Little Wenlock, on the other hand, he found that although the country rock immediately below the igneous body had been given a

TABLE 1.—CONTACT ROCKS FROM THE MIDLAND BASALTS. (AFTER EVERITT)

| Site | Position of Rock | Distance from Contact | Type of Rock | No. of discs | Azimuth A | Dip I | Circle of 95 per cent confidence radius $\alpha$ | Intensity $J \times 10^6$ gauss/gm. |
|---|---|---|---|---|---|---|---|---|
| Kinlet | Bottom contact | 1 ft. | Sandstone | 6 | 202° E. of N. | 17° D. | 4·4° | 27 |
| | Top contact | 6 in. | ,, | 6 | 210° | 7° D. | 2·2° | 36 |
| | Country | ? | ,, | 12 | 187° | 0° | 2·6° | 0·77 |
| Shatterford | Bottom contact | 6 in. | Shale | 6 | 173° | 17° U. | 2·7° | 14·0 |
| | Top contact | 6 in. | ,, | 6 | 160° | 15° U. | 5·3° | 5·4 |
| Pouk Hill | Top contact | 5 ft. | Shaly coal | 6 | 194° | 13° D. | 2·9° | 0·5 |
| | Above top contact | 15 ft. | Dark mud-stone | 12 | 205° | 1° D. | 24° | 48 |
| Little Wenlock | Bottom contact | 6 in. | Sandstone | 2 | 10° | 13° U. | Small | 1·2 |
| | | 1 ft. | Limestone | 12 | 349° | 1° U. | 14° | 0·4 |
| | Top contact | 1 ft. | Sandstone | 7 | 334° | 72° D. | 45° | 0·45 |
| | Country | ? | ,, | 6 | 329° | 57° D. | 20° | 0·2 |




(A) Variation of magnetic direction of a partially heated rock 1·4 metres from the contact.




(B) Variation of magnetic intensity of partially heated rocks.

TEXT-FIG. 6.—Thermal demagnetization of a lava partially heated during intrusion of a nearby dyke. (Everitt, 1959).

thermoremanence, that above had not, indicating that the body is a lava. Thus Everitt has been able to verify and strengthen Marshall's interpretation.

At an intermediate distance from the contact of an intrusion such as a dyke or a sill, where the country rock had been heated to some temperature less than the curie point, the remanent magnetization should retain a memory of the heating. The maximum temperature involved could be established by reheating specimens in the laboratory and noting the temperature at which the P.T.R.M. is destroyed. A series of samples collected at various distances from the contact would provide a curve of the variation of the maximum temperature acquired. By extrapolation, or by comparison with theoretical curves that have been worked out by Jaeger (1957), it would be possible to infer the temperature of the intruding magma. So far the full experiment has not been carried out, but Everitt (1959) has successfully reheated some partially heated specimens in the laboratory. Text-fig. 6 shows the result of an experiment carried out on a specimen taken from a Devonian lava 1·4 m. from the contact with an intruding Tertiary dyke. At around 500° C. the Tertiary P.T.R.M. was destroyed, leaving the remains of the original Devonian magnetization. The chief difficulty in performing such experiments is that chemical changes are apt to take place as the specimen is heated in the laboratory, and the effect of these often masks those under investigation.

## Summary

The examples described serve to illustrate some of the ways in which rock magnetism may be of practical use to the geologist. In most cases they are a means of correlation and depend only on the need to demonstrate magnetic stability and that magnetic directions are characteristic of a particular age or formation. It should be stressed that no assumptions are required regarding the nature of the past earth's field as a whole nor regarding hypotheses of possible polar migration and continental drift. Although rock magnetism is being used in the study of such hypotheses it can be used for geological purposes of a more humble nature in ways which are quite independent of them, and such uses have wide application. Although the examples given here are concerned with igneous rocks the methods are also applicable to sediments which contain some ferromagnetic material.

The geological use of the inherent properties of the magnetism of rocks is as yet hardly touched upon. The study of the contact metamorphism associated with a minor intrusion is merely a preliminary to the study of the thermal aureoles about major intrusive bodies. Later, perhaps, in conjunction with studies already under way of the effects of stresses on magnetic direction and susceptibility, rock magnetism may

help unravel the complicated histories of regionally metamorphosed rocks and add to the knowledge of such processes as granitization.

It is apparent that present knowledge is sufficient for rock magnetism to be used in the solution of a number of geological problems, but further and more detailed information is required if it is to become a precise and reliable tool.

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# The Relation Between the Quartz Fabric and Fold Movements in a Folded Calc-Silicate Rock from the Moines of Scotland

By T. J. Dhonau

Abstract

Petrofabric analyses of quartz *c*-axes have been made at four locations round a fold in a calc-silicate rock. The movement direction during the folding is deduced from the deformation of an older lineation. Two fabrics are present: the earlier, associated with the deformed lineation, is being replaced by a fabric associated with the formation of the fold. It is shown how the symmetry of the later fabric is related not to the axis of the fold but to the direction of the movement that produced the fold; and how the quartz *c*-axes lie on a conical surface, giving a small circle on a stereogram plot.

This study was made on a specimen from a thick band of Moine calc-silicate rock in the Five Sisters of Kintail, Ross-shire. The band outcrops in the bed of Allt a' Chruinn, east of the summit of Sgùrr an t-Searraich (NG 955193). It shows the effect of at least two periods of deformation, having preserved early structures on every scale that have almost vanished in the neighbouring siliceous and pelitic Moine rocks.

The specimen studied is a complete antiformal fold, shown at Text-fig. 1. The rock is banded, some bands being rich in actinolite, others in diopside; a few contain mainly. Quartz and plagioclase (oligolase) with a little microcline. Quartz and feldspar form the matrix in the layers rich in actinolite and diopside.

On the folded foliation surface is a lineation oblique to the axis of the fold and making a variable angle with the axis. This lineation lies on the intersection of a plane with the folded foliation surface, a feature shown by Weiss (1959) and Ramsay (1960) to be characteristic of shear folding. Inasmuch as the fold was formed during the last period of major deformation in the district the lineation must have been impressed on the rock during a previous period of deformation and then distorted. Near this exposure, folds of an earlier period which have been refolded during the later period support this interpretation. Therefore, in accordance with the work of Ramsay (1960), the intersection between the plane now containing the lineation and the axial plane of the fold is considered to be a movement direction during the folding—the *a* direction.

Within the body of the rock another lineation, which lies in a plane parallel to the axial plane of the fold, and which makes an angle of about 30° to the axial direction, is found. This lineation is formed by the alignment of elongated grains of actinolite.

The orientations of the axial plane, of the fold axis, and of the plane containing the old lineation were measured, and are shown on each stereographic plot.

Four analyses of the orientation of quartz *c*-axes were made, and the data obtained were plotted on the lower hemisphere of a Schmidt net; the scatter-diagrams so produced were contoured for density of points and these diagrams are presented in Text-fig. 1. The plane of the stereogram is that of the cut surface of the specimen; this surface was horizontal, and the orientation of the nets and of the specimen is shown by the compass points. The cut surface originally faced downwards, and is here drawn as though viewed from below.

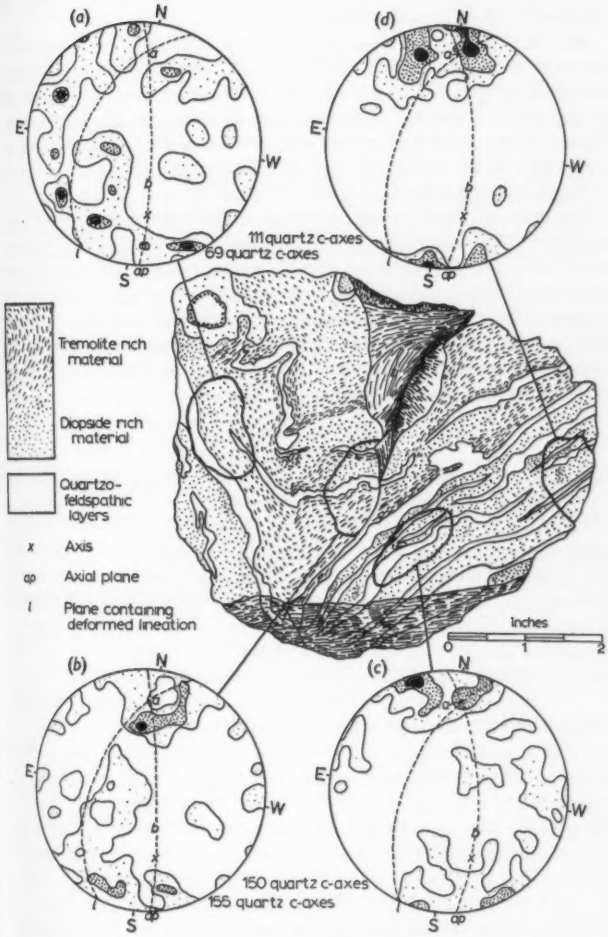
The fabric changes around the fold. In the diagram (a), where 69 quartz *c*-axes were measured (all the grains in the slide), there is a broken great circle, very diffuse, not far off the plane of the old lineation; there is also a broken but better developed small circle whose centre is near the intersection of the plane of the old lineation and the axial plane of the fold, the intersection considered by Ramsay to be the *a* direction of the fold movements.

In diagram (b) (155 *c*-axes), there is a rudimentary great circle, much where it was in the previous diagram, though even less well developed. The small circle around the *a* direction, however, is more emphatic, and the axes show less scatter than those of (a).

Diagram (c) (150 quartz *c*-axes), does not show the great circle which lies near the deformed lineation; the small circle is more strongly developed than before. A few axes are scattered, lying on neither of the girdles.

However, diagram (d) shows very few scattered axes, nor is there a great circle. Nearly all the 111 quartz *c*-axes lie on the conical surface plotted as a great circle, with the *a* direction near its centre. The *a* direction is sufficiently near the centre of the small circle for the discrepancy to be experimental error, so the small circle probably is centred on the *a* direction.

There are three noteworthy features of these diagrams: first, that the great circle seems to be associated with the plane containing the deformed lineation; second, that the great circle vanishes as one progresses round the fold; and third, that the small circle is associated with the movements during which the fold was formed. A possible explanation of these features is that an old fabric, which formed an *ac* girdle around the lineation which was later deformed, is being progressively attentuated and distorted before it is completely destroyed. In this series of diagrams, the diffuse great circle at diagram (a) is replaced by an intense small circle associated with the fold's formation, whose radius is about 20°, and which is centred on the direction of movement. As there is a complete girdle, no regularly oriented glide plane is




TEXT-FIG. 1.—Specimen of deformed Moine rock from Ross-shire, and four stereograms (a)–(d) derived from it, as described in the text. Contours are at 1%, 3% and 5% area.

operating; such a small circle could arise with the edge between rhombohedral or pyramidal faces as glide line.

Whereas these quartz *c*-axes are nearly horizontal, the actinolite grains forming the lineation within the rock mentioned above lie almost vertically, normal to *a*, and in the axial plane; they lie in the *b* direction of the fold. There are thus two mineral directions at right angles, formed by a single set of movements, one associated with the quartz, the other with the actinolite, due, no doubt, to the different behaviour of the two minerals during movement under stress.

Examples of small circles have not often been reported, though Fairbairn (1949) mentions them. The fabric here described is particularly interesting, for the small circle is associated with the movements during the deformation when the fold was formed and not symmetrically disposed with respect to the fold axis. It seems that although both the fold and the fabric are expressions of the movements, there is not necessarily any relation between the symmetry of the fabric and the shape of the fold; here they are quite distinct.

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# The Deformation of Confined, Incompetent Layers in Folding

By Emyr Williams

Abstract

Shearing stresses acting on an incompetent layer confined between competent beds during its deformation in the development of folds, differ from those commonly accepted as the cause of a number of minor structures associated with the deformation. It is suggested that the effect of bedding drag is negligible, and that the structures described as slaty cleavage, fracture cleavage, and drag folds can be explained by the laminar flow of the incompetent material whilst accommodating itself to the interspaces determined by the more competent layers.

## I. Behaviour of Beds in Folding

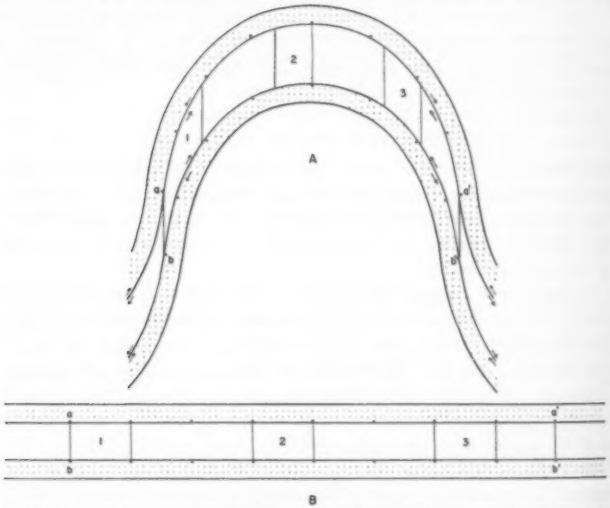
The terms " competent " and " incompetent " refer to the structural characters of rocks, and are used in such a way that a weak rock which has undergone considerable deformation is called " incompetent ". The terms are relative, and are invaluable in discussing the behaviour of beds during folding.

It is generally recognized that the folds commonly encountered in geology involve rocks of varied competence, and that the fold patterns are determined by the most competent beds. During the flexing of competent layers, the thicknesses of which remain approximately unchanged normal to the bedding plane, any interbedded incompetent material is believed to behave more or less passively, and to adjust itself to the interspaces (Leith, 1913, p. 114 ; Willis, 1923, p. 25 ; Hills, 1953, p. 90). The permanent deformation of the incompetent horizons is considered to be by plastic flow, where the material yields without rupturing (Mead, 1940, p. 1012).

It has frequently been stated that where competent and incompetent strata are interbedded, the movements of the former relative to each other during folding subject the incompetent beds to shearing stresses, which are considered to play an important part in the deformation. These shearing stresses are incorrectly believed to be similar to those developed in the concentric folding of competent strata, where upper beds tend to ride over lower ones towards the anticlinal crests, and a number of minor structures associated with the deformation of the incompetent material have been attributed to them (Leith, 1913, p. 114 ; Wilson, 1946 ; Nevin, 1949, pp. 73, 166 ; Hills, 1953, p. 97).

The sense of bedding-plane shearing can only be deduced by examining the movements of strata relative to their adjacent beds. Assuming that the layers of a sequence of competent and incompetent rocks were originally of uniform thickness when they were horizontally disposed, a consideration of the geometry of such beds in similar folds

indicates that material of the incompetent layers migrates from the limbs towards the crests and troughs during folding (see Hills, 1953, Fig. 49A). Text-fig. 1A is the profile of an anticline of a similar fold that may develop from a sequence of competent and incompetent beds, the original thicknesses of which are shown in Text-fig. 1B. It is evident that the relative movements of the material of the incompetent and competent layers during folding causes a shearing force on the top and




TEXT-FIG. 1.—Profile of anticline of a similar fold (A) developed from sequence of competent (dotted) and incompetent beds, of original approximate thicknesses shown in B. No allowance made for either migration of incompetent material in direction parallel to fold axis, or any volume change. Lines ab and a'b' are approximate boundaries on one side of which material moves towards crest, and on other side to adjacent troughs. Comparison of points of reference in competent layers (large dots), and areas 1, 2, and 3 in incompetent bed, of diagrams A and B, gives relative movements. Shearing stresses are indicated by arrows. Vertical margins of areas 1, 2, and 3 in diagram A may not be strictly true in direction. However, shearing stresses remain as given for all reasonable deviations of margins from positions indicated.

bottom of the incompetent bed in the anticline directed downward away from the apex of the anticline. The direction of the shearing force developed on the top of the incompetent bed is, therefore, opposite to that commonly accepted. Similarly, shearing stresses indicated by the relative movements of the material of the incompetent and competent layers in disharmonic folding differ from those usually implied in the text-books. In the case of an incompetent bed thickening in the limbs and thinning in the crests and troughs, the shearing force on the top and bottom of the bed in an anticline is upward towards the anticlinal apex.

## II. Minor Structures of Confined, Incompetent Layers in Folds

The significance of bedding-plane shearing stresses, and a more detailed movement pattern within incompetent material confined between competent beds during folding, can be determined by considering the minor structures that may develop within the incompetent layers. In the following account only the common structures are described, and the mechanism of their formation discussed.

(1) *Slaty cleavage.*

Slaty cleavage is referred to as flow cleavage, which has been defined as "a capacity of some rocks to part along parallel surfaces, not necessarily planes. These surfaces are determined by the parallel dimensional arrangement of the mineral constituents, that is, by the mutual parallelism of the greatest, mean, and least dimensional axes of the mineral particles making up the rock mass. They may also be determined by the parallelism of the mineral cleavages of the constituent particles" (Leith, 1913, p. 76). It is generally accepted that the parallel arrangement of the minerals is brought about by recrystallization, granulation, and rotation of original particles (Leith, 1913, pp. 79–84; Harker, 1939, pp. 154–155; Collette, 1958, pp. 123–127).

Slaty cleavage is more or less parallel to the axial planes of folds, and has been shown in a number of studies of shear folds to have developed in the direction of elongation of the rock mass (Sharpe, 1847; Cloos, 1947).

Many authors consider that slaty cleavage may result from forces which can be resolved into a single applied force and a turning couple, which in the case of confined, incompetent layers is considered to be generated by bedding-plane slip of the sense that is present in concentric folding (Wilson, 1946, p. 281). However, it has been shown above that the bedding-plane shearing developed during the folding of an interbedded sequence of competent and incompetent layers differs from that in concentric folding, and it differs in such a way that the effect of bedding-plane drag must be insignificant in the development of slaty cleavage.

Cloos (1947) concluded from his examination of the deformation of oölites in the South Mountain Fold, Maryland, that the oöids, elongated in planes parallel to the slaty cleavage in the shear fold, must have been distorted in a plastic condition, and that the geometry of deformation could be explained by laminar flow in sub-parallel planes, in which the cleavage developed.

(2) *Fracture cleavage.*

Fracture cleavage "consists of closely spaced, parallel . . . ruptures . . . in no way dependent on the orientation of constituent minerals"

(Mead, 1940, pp. 1009–1010). In the field, fracture cleavage, which is sub-parallel to the axial planes of folds, may vary from a very finely fissile structure, commonly found in the finer grained layers, to a close joint system, often encountered in coarse sandstone and limestone.

Many writers have stated that fracture cleavage is due to a rotational stress caused by the frictional drag between beds while they are being folded, and have discussed the mechanism of formation of the cleavage in terms of the strain ellipsoid theory of Becker (Leith, 1913, pp. 23–25; Wilson, 1946, pp. 273–284; Nevin, 1949, p. 167). The application of this theory is based on the assumption that the rock was appreciably deformed before rupture occurred, and that the orientation of the cleavage depends on the orientation in the hypothetical strain ellipsoid of the planes of no distortion, the position of which is dependent, in the main, on the rotational stress caused by bedding-plane slip of the sense occurring in concentric folding.

A number of criticisms can be made of this suggested mechanism. Hills (1940, p. 101) has observed that fracture cleavage is frequently best developed in the neighbourhood of the fold axes, where least bedding-plane slip is to be expected. Again, a consideration of the geometry of folds involving both competent and incompetent strata, indicates that the strain ellipsoid theory cannot be applied, for the bedding slip differs from that necessary to account for the formation of fracture cleavage.

Hills (1953, pp. 105–106) has drawn attention to the belief of White (1949), that planes which appear to be simple shearing planes may be analogous to the flow planes found in rolled, or extruded metals, in wet clay, or in liquids. As in the case of slaty cleavage, fracture cleavage may, therefore, be parallel to the elongation of the rock mass, which was first suggested by Sorby (1853). De Sitter (1956, pp. 214–215) regards much of fracture cleavage, similarly, to be in the direction of mass elongation, and he considers this type of cleavage to be shearing planes resulting from the influence of elastic strain on the process of plastic deformation.

It is evident in a number of cases that fracture cleavage is genetically related to slaty cleavage, for there are many examples known where one grades into the other (Leith, 1913, p. 62; Wilson, 1946, p. 281; Hills, 1953, pp. 108–109).

Recently, Collette (1958) presented a theory of schistosity based upon the elastic anisotropy of the elementary crystal components, and recrystallization according to Riecke's principle. Schistosity, "when new-formed mica appears macroscopically on the planes of fissility" and fracture cleavage, "when this is not the case" (p. 121), are believed to develop in flow planes during plastic deformation, and they are considered to be different stages of one and the same process.

However, Collette maintains that fracture cleavage occurs only after orientation of mica of microscopic dimensions has made the material anisotropic. Nevertheless, by Mead's definition it appears well established that fracture cleavage is not dependent on mineral orientation.

(3) *Drag folds.*

Drag folds are minor folds and contortions commonly occurring in confined, incompetent strata in major folds. Only those drag folds whose axial planes and plunges agree with the major structure are considered. The secondary folds that have resulted from movements other than those produced by the major fold, and convolute folds that may have been formed by intrastratal flow of liquefied layers during the development of the major fold (Williams, 1960), are excluded.

Drag folds have been attributed to the shearing of the layer in which they occur between more competent beds, and the sense of the shearing is considered to be similar to that developed in concentric folding (Leith, 1913, p. 114 ; Wilson, 1946, p. 291 ; Nevin, 1949, p. 73 ; Hills, 1953, p. 97). However, these minor folds are often best developed near the crests and troughs of the major folds (Nevin, 1949, p. 73), where bedding-plane slip is at a minimum, and differs from that required.

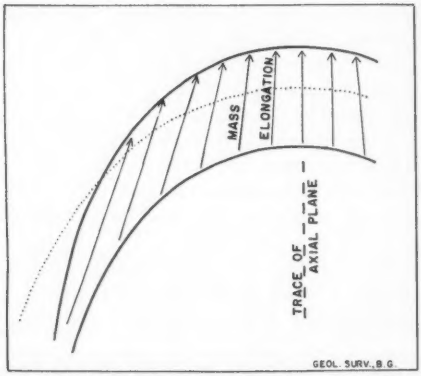
Numerous examples are to be found described in the literature, of drag folds of bedding laminations developed in confined, incompetent layers, where the folds are due to differential movement along cleavage, and along planes paralleling cleavage (Willis, 1923, figs. 30–31 ; de Sitter, 1956, figs. 63–66, pp. 225–226). Cloos (1947) demonstrated a genetic relationship between the development of drag folds and cleavage in his investigation of the deformation of oölites in a shear fold. He observed that the bedding-plane crenulations, which first appear at about a 20 per cent deformation of the oöids, become more intense with increasing distortion of the components of the oölites, and that cleavage, which appears at a 30 per cent deformation, becomes more dominant. The development of both the bedding-plane folds and the cleavage is in the direction of mass elongation, as indicated by the deformed oöids. It appears likely, therefore, that much of drag folding is of the type described by Carey (1954, pp. 91–98), and probably all can be explained by differential movements along laminar flow planes, in which cleavage may develop.

## III. Conclusions

In the development of folds involving competent and incompetent layers, the fold pattern appears to be determined by the competent layers, whilst the material of incompetent beds accommodates itself to the interspaces by flow, which according to a study of the geometry

of deformation, and a consideration of minor structures, can be but little influenced by bedding-plane drag.

It is probable that in the early stages of folding, incompetent material, often mudstone, flows by intergranular movements, but as folding proceeds movement by recrystallization, crushing, and minute fracturing becomes important (Mead, 1940, pp. 1012–1016). During




TEXT-FIG. 2.—Portion of profile of anticline in which continuous lines indicate top and bottom of bed which behaved incompetently during similar folding of confining competent layers. The deformed bed, originally of uniform thickness, is compared with layer of same thickness which behaved competently in folding: the bottom of layer is shared with that of deformed bed, whilst top is indicated by dotted line. From comparison it is evident that mass elongation within confined, incompetent layer during deformation is approximately in direction shown by arrows.

the deformation of the confined, incompetent layer there is orientation of the component particles, and minerals of platy habit develop with their longest dimensions in the direction of mass elongation, which is approximately in the direction shown by the arrows in Text-fig. 2. Where good orientation is achieved slaty cleavage results. The geometry of the development of slaty cleavage has been adequately accounted for in terms of laminar flow, where the planes of flow are in the direction of mass elongation, and are more or less parallel to the axial plane of the fold.

In laminar flow a shear strain exists between each lamina. It is, therefore, suggested that if the deformation of incompetent material takes place at a rate which does not allow internal rearrangements, then fracturing results in the planes of flow. In this way fracture cleavage could be formed.

Differences in the degree of the ease with which internal rearrangements take place within layers, or within bedding laminations of a layer, may result in the gradation between slaty cleavage and fracture cleavage.

Differential movement between flow laminae accounts for the formation of drag folds within an incompetent layer. Such differential movement may take place before, or after, orientation of the components within the material is achieved, or along fractures if the material has failed by rupture in the flow planes.

It should be pointed out that the direction of flow planes can vary according to the degree of incompetence of the material, and the character and degree of flexuring of the confining competent beds. Thus the dip of the cleavages and the axial planes of drag folds, with respect to the axial plane of the major fold, may vary considerably.

Again, a bed that behaves incompetently during the early stages of folding may become as competent as those that confined it. In such cases forced deformation of any minor structures, that have formed within the layer, could result from shearing of the sense generated in concentric folding.

## IV. Acknowledgements

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# A Note on Cleavages in Crenulated Rocks

By M. J. Rickard

Abstract

Published descriptions and new field observations in Ireland and Canada confirm that strain-slip cleavage should not be classified with fracture cleavage. The term *crenulation cleavage* is proposed to replace the generic one of strain-slip cleavage. Crenulation cleavage differs essentially from slaty and fracture cleavage in that it only develops in laminated rocks and is consequently nearly always a secondary structure. Intense metamorphism converts crenulation cleavage into a new schistosity or cleavage with characteristics similar to slaty cleavage. Two fundamentally different types are recognized and it is suggested that crenulation cleavage is a valuable indicator of metamorphic conditions during the various phases of deformation.

## I. Introduction

Cleavages associated with small-scale crenulations are frequently present in metamorphic rocks, particularly in those areas where repeated folding has occurred. In much of the literature this phenomenon termed strain-slip cleavage or, in America, slip-cleavage is usually considered to be a variety of " fracture cleavage ". Extensive references are given by Wilson (1946) and King (1956).

Leith classified strain-slip cleavage with fracture cleavage despite the fact that it is marked, in many cases, by a parallel arrangement of minerals which is contrary to his ruling that " Fracture cleavage ... is independent of a parallel arrangement of the mineral constituents " (Leith, 1905, p. 12). Leith's view has been widely accepted and is still in vogue today. Mead (1940), however, interprets strain-slip cleavage as a shear phenomenon developed on planes at 45° to the compressive stress, and thus to the axial planes of minor folds. Although such conjugate cleavage sets have been described (Muff, 1909 and Agron, 1950) they are not commonly developed and represent special cases. Studies in the Appalachians (White, 1949 and Brace, 1953) have indicated that the characteristics of strain-slip cleavage relate it more closely to slaty (flow) cleavage than to true fracture cleavage.

During detailed investigations of Dalradian rocks in North-west Donegal, Ireland, and of Paleozoic rocks in the Appalachians in southern Canada, the writer found the previous terminology and treatment to be unsatisfactory. As many of the characteristic features that may be summarized from the literature are present in the Errigal area Donegal, the cleavage structures found there are described as a basis for discussion of cleavage in complexly folded schistose rocks, and its significance with regard to interpretation of their metamorphic and tectonic history.

The writer would like to thank Dr. J. N. Knill and others at Imperial College and at McGill University who discussed and criticized the ideas presented in this paper. Field work in Ireland was supported by a D.S.I.R. grant and in Canada by a National Research Council Fellowship and these are gratefully acknowledged.

## II. Nomenclature

The term strain-slip cleavage was first introduced by Bonney (1886, p. 95) as a translation of Heim's (1883) term " ausweichungsclivage ". It was applied to a " cleavage foliation produced by strain and rupture of flexures " (Bonney, 1886, p. 98). Various other terms have been applied to this structure namely; close-joint cleavage, false cleavage, shear cleavage, schistosity shear, slip cleavage, and fracture cleavage. The terms strain-slip and slip cleavage have genetic implications that are particularly bad because the structure can almost certainly originate in several different ways. A completely descriptive name is desirable and it is proposed here to introduce the term *crenulation cleavage* (this term appears in German literature as " runnzelsclivage " Hoeppner, 1956).

The cleavage terminology used in this paper is based mainly on Leith's established definitions, applied in the *strictest* sense, as follows :—

(i) *Slaty cleavage*, *flow cleavage*, *and schistosity* are regarded as synonomous terms for " the cleavage dependent on the parallel arrangement of the mineral constituents of the rock " (Leith, 1905, p. 23).

(ii) *Fracture cleavage* is a distinct break or rupture cutting a rock with no regard for its mineral arrangement. For a discussion of slaty and fracture cleavage the reader is referred to Wilson (1946).

(iii) *Crenulation cleavage* is used simply to designate cleavage planes, whether micaceous layers or sharp breaks, which are separated by thin slices of rock containing a crenulated cross-lamination.

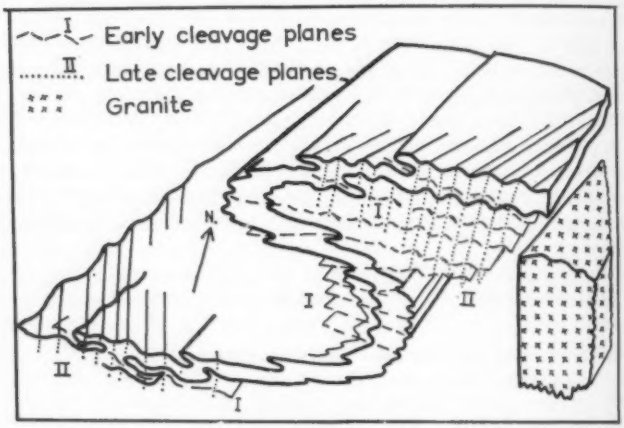
In describing folds, *major* is used for a fold large enough to show on a map, *minor* for a fold of outcrop size, and *micro* for crenulations between cleavage planes (from a fraction of an inch up to about 1 inch across).

## III. Cleavage Structures in the Errigal area, Co. Donegal, Ireland

The Errigal area is situated in North-west Donegal lying between the Main Donegal Granite and the Bloody Foreland Granite (Pitcher and Cheesman, 1954).

The general structure is that of a major recumbent fold defined by a

thick massive quartzite formation (Text-fig. 1). A bedding schistosity can be traced around this fold and is often crumpled into minor folds with a rounded S-shaped profile. Both slaty and crenulation cleavages are found in the axial planes of these minor folds, and apparently both types are related to the formation of the recumbent fold. Superimposed upon these early structures, there are two zones of later minor folds with steep axial-plane crenulation cleavages, which are closely related to the emplacement of granite intrusions. (Text-fig. 1).




TEXT-FIG. 1.—Generalized structure of the Errigal area, showing the early flat cleavage planes contorted by two sets of later folds with steep axial planes. The axes of the late folds trend in different directions from that of the major fold structures.

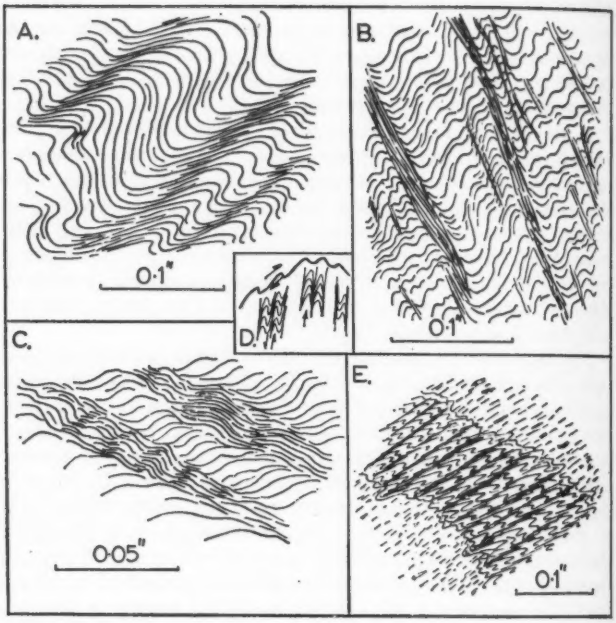
A complete sequence in the development of crenulation cleavage can be observed in the pelitic groups. The first stage is marked by a corrugation of the regional bedding schistosity ; then, as folding proceeds, the micro-flexures become more strongly developed, producing a marked " herringbone " structure. In many cases these flexures are seen to behave as miniature drag folds with respect to minor folds in the more competent beds. Thinning of the long limbs of the micro-flexures occurs as the quartzose material migrates to the crests and short limbs of the flexures. In this way a micaceous banding is produced as an incipient cleavage parallel to the fold axial planes. This banding consists of narrow zones in which the individual micas, although parallel to each other, are set across the length of the band (Text-fig. 2A). The micas defining the micro-folds between these incipient cleavage planes are, in different situations, curved, bent, or even

broken, and the micro-fold style varies from gently rounded to sharply angular. More intense deformation and folding causes further rotation of the micas into the axial plane direction. This process together with limited re-crystallization forms a fine cleavage banding along which differential slip takes place (Text-fig. 2B).

In the limbs of the major and minor folds the micas are frequently aligned parallel to the axial planes, providing a schistose arrangement resembling slaty cleavage. Crenulation cleavages are often preserved in the cores of the minor folds and can be traced into this slaty cleavage, suggesting that the latter has been derived from or developed at the same time as the crenulation type. Also, some banded pelites indicate a transitional stage, where a closely spaced axial-plane crenulation cleavage in micaceous bands can be traced as a slaty cleavage through semipelitic bands (Text-fig. 2E).

The later superimposed cleavages show a similar sequence of formation, although in this case they are initiated as flexures on the earlier crenulation and slaty cleavages (Text-fig. 2C). They, too, can be traced into a nearly perfect schistosity towards the Bloody Foreland Granite. The concentration of mica in the cleavage planes becomes strongly emphasized by metamorphic differentiation, so that, in places, a regular close banding across the bedding planes results. White (1949, p. 590) describes a somewhat similar transition of a crenulation cleavage into a new schistosity, from the staurolite zone of a granite thermal aureole ; he reports that " Study of the rock from the zone of transition suggests that the later schistosity has formed in part by a mechanical rotation and smearing out of earlier schistosity, and in part by the growth of new mica flakes orientated parallel to the later cleavage ". Such a mechanism could well account for the cleavage banding and transition from crenulation to slaty cleavage mentioned above.

The minor folds associated with these crenulation cleavages, more especially the superimposed ones away from the granite, commonly have an angular zig-zag profile. The cleavages in the axial planes of these folds are often distinct breaks and the thin laminae between them are only slightly flexed with little, if any, rotation of mica into the cleavage direction. Occasionally, however, films of new mica lie in the cleavage planes. It is probably this type of cleavage that has caused most of the confusion in the nomenclature. According to strict definition they are planes of fracture cleavage for they are independent of a mineral arrangement. Nevertheless, they are genetically related to the associated crenulation cleavages and, like them, are oriented parallel to the fold axial planes. Several examples of cleavage showing transitional features between this fracture type and micaceous crenulation cleavage can be seen in the Errigal area.




TEXT-FIG. 2.—Crenulation cleavages.

(A) Folded bedding schistosity with development of an incipient crenulation cleavage.

(B) Crenulation cleavage—micas are completely rotated into discreet zones parallel to the axial planes of the micro-folds, cleavage slip occurs.

(C) Flexured crenulation cleavage planes—an initial stage in the development of a second crenulation cleavage.

(D) Opposite senses of rotation indicated by micro drag-folds and cleavage plane slip.

(E) Banded pelitic rock, showing strong crenulation cleavage in micaceous bands and slaty cleavage in more quartzose bands.

All figures are drawn from photo-micrographs.

It is commonly assumed that the crenulation results from small fault movements along the cleavage planes. But emphasis must be placed on the fact that both the early and late cleavage planes in this area developed as a consequence of micro-folding and not vice-versa, for as White (1949, p. 590) says, " The crinkles are the independent part and the slip-cleavage the dependent part of the phenomena." It must also be noticed that generally the sense of rotation implied by the differential movement along the cleavage planes is in the opposite direction to that

of the micro-folds produced by interlayer slip on the limbs of controlling minor folds (Text-fig. 2D). In the fold cores cleavages are developed on both limbs of the micro-folds so that there is only extension in the axial plane.

## IV. The Significance of Crenulation Cleavage

The most important feature of crenulation cleavage is that it can only develop from a pre-existing schistosity. This infers an earlier deformational or structural history, and in nearly all the examples in the literature, crenulation cleavage is associated with secondary or superimposed folding. The initial schistosity is usually either a bedding cleavage or an earlier axial-plane cleavage.

The stress orientation during refolding is often the same as for the earlier deformation (see Wilson, 1946), but it can be completely different depending on the local or regional forces operating at the time. Thus, crenulation cleavages occur in the axial planes of cross-folds in the Appalachians and also in the Central Highlands of Scotland (Rast, 1958).

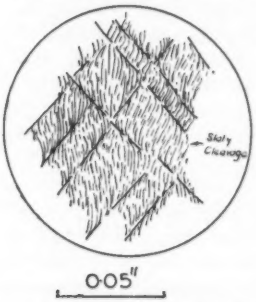
The presence of a crenulation cleavage cutting an earlier slaty cleavage indicates a change in metamorphic conditions. The presence of slaty cleavage changes the physical properties of a rock so that its behaviour during refolding will of necessity be different. However, when either the deformation or metamorphism is intense during refolding the crenulation cleavage can develop into a new schistosity. Knopf (1931) and Turner (1942) consider crenulation cleavage an intermediate stage in the " transposition " of a schistosity into a new direction. This process can operate any number of times and two or three superimposed crenulation cleavages are common. In many cases the obvious latest structure bears no relation to the earlier fold systems, a point of special importance in fabric studies.

The various types of crenulation cleavage are clearly determined by a delicate balance of different factors of which interlayer slip and metamorphic intensity are the most important. With low metamorphism interlayer slip is dominant and sharp zig-zag folds with fracture type cleavage planes develop. With higher metamorphism the strength of the individual laminae is reduced and recrystallization is the dominant factor. Micas concentrate in the axial planes and quartz migrates to micro-fold crests. Different rocks will naturally behave differently under given conditions and in the Errigal area associations of fracture, slaty, and crenulation cleavage can be seen in flaggy quartzites, semipelites, and pelites.

These observations are borne out in the Appalachians of southern Canada. There a late deformation folds an early bedding schistosity and crenulation cleavage into a major anticlinorium. Metamorphism

is most intense in the core of this fold and mica is commonly segregated into the late cleavage planes in this position. Further out where metamorphism is weaker, the late minor folds and crenulations are more angular, and the cleavage planes sharper with much less segregated mica.

A crenulation cleavage developed on sets of *conjugate planes* has been described by Muff (1909), Mead (1940), and Agron (1950) and recently from the Craignish area of the South-west Highlands of Scotland by Knill (1960). In these cases the crenulation cleavages develop as narrow zones of flexure crossing the early slaty cleavage at approximately 45° angles (Text-fig. 3). They differ from the axial




TEXT-FIG. 3.—Conjugate crenulation cleavage cutting an earlier slaty cleavage from Craignish area, Scotland (by J. N. Knill).

plane types in three respects—minor folds are rare, the sense of movement along each set is consistent throughout the area, and recrystallization is restricted to the cleavage zones (Knill, 1960).

The factors leading to the development of a conjugate crenulation cleavage rather than an axial plane type are not clear but from the meagre literature it seems that they only develop under low grade metamorphic conditions at a very late stage of tectonic activity. At Craignish a continuation of the original forces led to failure of the slates on conjugate shear planes symmetrically disposed about the early cleavage (Knill, 1960). Being laminated the rocks failed by flexure rather than simple rupture.

It is possible that these conjugate cleavages may be more widespread than has hitherto been appreciated ; Harker (1950, Fig. 68, p. 158) for example, figures them in rocks from Snowdonia and Cornwall. Sometimes only one set of a conjugate pair is developed (Knill, 1960), and in such cases it may be difficult to determine which type is developed. This would be especially true if, as in a case described by Richey (1948), the crenulation cleavage is at 45° to an earlier cleavage.

## V. Summary and Conclusions

The main features of crenulation cleavage from the writer's observations and the published literature are summarized below :—

(1) It is found only in finely laminated micaceous rocks.

(2) The necessary lamination is provided by an earlier cleavage, schistosity, or bedding schistosity.

(3) It is initiated by flexure rather than by microfaulting (there may be types in which the crenulations result from slip on the cleavage planes but such an origin has yet to be demonstrated).

(4) The shape of the micro-folds varies from slightly curved to sharply plicated; the individual micas may be bent, broken, or recrystallized into new straight elements.

(5) The cleavage varies from sharp fracture like breaks to micaceous types similar to slaty cleavage.

(6) A preferred mineral orientation develops along the cleavage planes and under certain conditions this may eventually result in the formation of a cleavage banding or a new schistosity.

(7) Interlayered beds of different composition, especially those without bedding schistosity, may show slaty or fracture cleavage.

(8) Crenulation cleavage commonly lies parallel to the axial planes of major and minor folds.

(9) A fundamentally different type occurs as two sets of planes developed simultaneously or approximately at the same time.

Crenulation cleavage should be distinguished carefully from fracture and slaty cleavage because of the special conditions necessary for its formation. Crenulation cleavage is a feature of anisotropic rocks, whereas slaty and fracture cleavage develop in essentially isotropic units. There must be a limiting thickness governing the isotropic behaviour of thin beds depending on the physical properties of the rocks and the environment. It can be demonstrated that where metamorphism destroys the anisotropic fabric (lamination) a new or "transposed" schistosity develops. Its dependence on an anisotropic fabric accounts for the common association of crenulation cleavage with refolding.

The several different cleavage varieties observed may have originated by different mechanisms (e.g. interlaminar slip or axial plane shear). Two main types are apparent, however, those developed in the axial planes of folds and those belonging to a conjugate set, usually without folds. The former develop under low to medium grade metamorphic conditions whereas deformation on conjugate planes of shear seems to occur only under very low grade conditions during the waning stages of

tectonic activity. Since they imply different tectonic conditions the differentiation of single and conjugate sets is important, but the distinction may be difficult where only one of a conjugate set is developed.

Study of the patterns of micro-folds and their associated cleavages gives valuable information regarding the physical conditions and the time association between metamorphism and tectonic activity for various phases of deformation.

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# Co-existing Pyroxenes in Igneous Assemblages : a Re-evaluation of the Existing Data on Tie-line Orientations

By G. M. Brown


## Abstract

Tie-line data are presented for twelve analysed igneous pyroxene pairs, from which it is demonstrated that a precise point of intersection, at or close to $Wo_{75}$ $En_{25}$ in the ternary diagram, does not exist. Eight tie-lines for pyroxene pairs from ultrabasic nodules in basalts are also considered. The tie-lines for metamorphic pyroxenes are compared with the igneous, and are shown to intersect over a similarly wide range, the average ($Ca_{82}$ $Mg_{18}$) being the same for both assemblages. It is shown that no conclusions can be drawn as to the genesis of the pyroxene assemblage, from the assumption of a characteristic tie-line intersection point. Optical determinations are considered, and shown to be of little practical value in this particular problem.


## Introduction

Twenty years ago Hess (1941) made the interesting observation that when the compositions of co-existing calcium-rich and calcium-poor pyroxenes from certain igneous rocks, as points plotted in the system $CaSiO_3$–$MgSiO_3$–$FeSiO_3$, were joined, then the projection of these joins intersected the $MgSiO_3$–$CaSiO_3$ side of the triangle at positions approximating to $Wo_{75}$ $En_{25}$ (op. cit. p. 585). Hess, in presenting the data, commented only very briefly upon the possible significance of this coincidence. Since then, however, this observation has been referred to in a great many petrological papers. The general tendency has been for the investigator to make optical determinations on the co-existing pyroxenes and, if the compositional tie-lines should project near to the $Wo_{75}$ point, draw the reader's attention to the fact. Presumably the cases (which may be many) where this coincidence is not found are not recorded, with the result that nowadays the petrologist is left with the impression that the exact position $Wo_{75}$ $En_{25}$ in the pyroxene composition triangle has a special significance.

Recently attention has again been focussed on this phenomenon by Muir and Tilley (1957 and 1958) and Wilson (1960), in their studies of co-existing pyroxene assemblages in metamorphic rocks. They were concerned, largely, with whether any significant difference existed between the direction of the tie-lines in metamorphic and in igneous assemblages ; for the purpose of comparison it was assumed in each case that the tie-lines for igneous pyroxenes intersected the Wo-En side of the triangle at $Wo_{75}$ $En_{25}$. Muir and Tilley (1958) reached the conclusion that no significant difference existed, whereas Wilson (1960)

concluded that differences did exist—not only between igneous and metamorphic assemblages, but also between metamorphic assemblages with differing genetic histories.

The present writer, when working on the co-existing calcium-rich and calcium-poor pyroxenes of the Skaergaard intrusion, naturally became interested in the suggestion made by Hess, particularly because the Skaergaard results (Brown, 1957, Text-fig. 2) provided four tie-lines based on chemical analyses of pure concentrates of each pyroxene of the four pairs. These tie-lines did not show a tendency to project to a single point, and disillusionment as to its significance was increased when tie-lines were plotted for all available analyses made on co-existing pyroxenes from igneous rocks. All these tie-lines were plotted (op. cit. Text-fig. 3), but their projections on the Ca–Mg join were not discussed. The author's reasons for neglecting discussion, at the time, were that the disagreement could easily be checked by the interested reader, the original suggestion of an exact point of projection by Hess was so tentative that it could hardly have been taken very seriously, and, perhaps more important, the exact position of the point seemed to have no real significance on, say, crystal chemistry or petrogenetic grounds. With reference to the last remark, it was felt that the real contribution made by Hess was proof that the projection was on the Ca–Mg, rather than the Ca–Fe join, and that, therefore, the ratio of Fe to Mg was always higher in the calcium-poor than in the co-existing calcium-rich phase. Also it looked as though the projections did not tend to lie around the apex, which would have implied that the Mg:Fe ratio was identical in each member of the pair, and that measurement errors resulted in a spread around the apex. To go further, and establish an actual point of some constancy on the Ca–Mg join, seemed hardly justifiable on the basis of available evidence.

Now, however, the picture is changing. Recent authors are accepting Hess's point at $Wo_{75}$ on the Ca–Mg join (referred to henceforth as " the $Wo_{75}$ position ") as a characteristic property of *igneous* pyroxene assemblages, and comparing it with metamorphic assemblages without further question. The object of the present paper is to question the validity of that accepted property of igneous assemblages, before too many conclusions are drawn through comparison. Already Wilson (1960) is taking the igneous pyroxene assemblages as evidence of a mineral association produced by equilibrium conditions, and from there suggesting that the nearer the projection of the metamorphic pyroxene tie-lines approach the $Wo_{75}$ position, the closer they are to having attained equilibrium between each other and, presumably, the other minerals in the rock. Such problems of metamorphic equilibrium are not the subject of this paper, only the problem of whether certain criteria can be classed as valid in the argument.

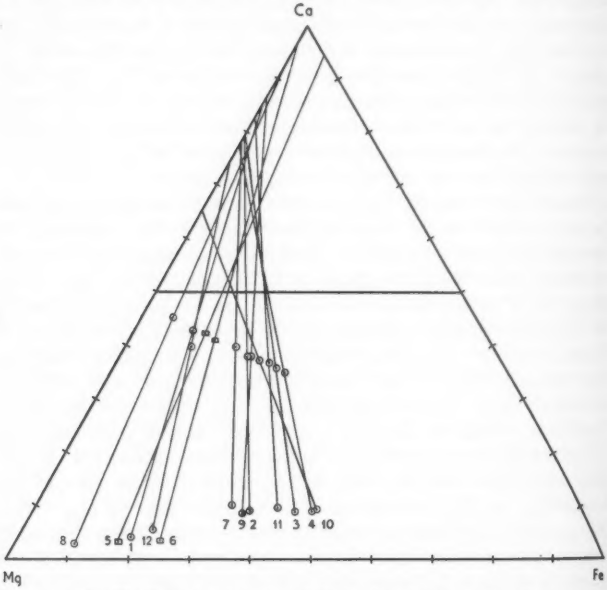
## Tie-lines for Igneous Pyroxenes

(*a*) *The data of Hess* (1941).—When Hess plotted the co-existing pyroxenes from igneous rocks he had available only a few chemical analyses of pure concentrates, together with some optical data. In order to plot the tie-lines used in his figure (1941, Text-fig. 11) he had chemical data for only *two* tie-lines. By reference to his Text-fig. 10, it would seem that one pair was from the Stillwater, and one from the Bushveld complex. Although the compositions of the Bushveld pair are not given, the Stillwater pair seems to be the one from 4,750 m. given in his Table 2, intersecting the Wo–En line at $Wo_{81}$. The intersection of two tie-lines, with an angle of deviation of only 11°, would be unlikely to provide an intersection point acceptable and reproducible in petrological literature for twenty years, yet such is the case, with reference to chemical information on the subject.

In addition, however, Hess plotted a further four tie-lines between pairs in which one or both members had not been analysed, but determined by optical methods. Such indirect methods are frequently necessary, but should be treated with caution. Thus, the optical properties of pyroxenes can only be of value, for subsequent plotting on a chemical ($CaSiO_3$–$MgSiO_3$–$FeSiO_3$) variation diagram, if used in conjunction with accurate optical/chemical correlation diagrams. Such diagrams were not produced by Hess until 1949 and 1952, and have since been modified slightly (e.g., Muir, 1951 ; Brown, 1957 ; Carmichael, 1960). In particular, attention has been drawn by Hess (1949) and the later investigators to the problems involved in correlation, particularly with reference to the effect of minor cations and sub-solidus exsolution (Brown, 1957 ; Muir and Tilley, 1958 ; Wilson, 1960). It is believed doubtful, therefore, that in 1941 a chemical composition could be obtained for a pyroxene, through optical determinations alone, which could be used with an accuracy comparable to a good (i.e., well charge-balanced) chemical analysis. In connexion with the chemical analysis it is worth emphasizing, at this stage, that contamination of one pyroxene sample by some of the other pyroxene in the same rock would not matter, for the purpose of drawing a tie-line, as the effect would be merely to narrow the gap between the two plotted points, rather than to affect the direction of the tie-line.

If, despite these reservations, the petrologist is prepared to accept all six tie-lines drawn by Hess, then the point of intersection lies between about $Wo_{71}$ and $Wo_{81}$ and averages at about $Wo_{75}$. On chemical grounds alone, however, the suggestion of a point of intersection at $Wo_{75}$ is based on only two tie-lines and cannot be said to be proven. In any case, whether all six or only two are accepted, they must now be viewed in relation to the tie-lines subsequently obtained for other pyroxene pairs obtained from the chemical analysis of each component.

(*b*) *Data obtained since 1941.*—Igneous rocks containing two pyroxenes in equilibrium, always a calcium-rich (augite) and a calcium-poor (orthopyroxene or pigeonite) variety, may be either of two types. Firstly, they may be volcanic rocks containing phenocrysts of each type, but these are rare for one type usually occurs in the groundmass, or may be xenocrystic or a phenocryst (or microphenocryst) of a different




TEXT-FIG. 1—Tie-lines joining co-existing, chemically analysed pyroxenes from igneous rocks. Lines 5 and 6 join pyroxenes from adjacent, rather than the same, rocks.

generation from the other. Secondly, they may occur in slowly-cooled basic rocks lying in the composition range in which two pyroxenes may crystallize almost simultaneously from the magma: a restricted field of crystallization which passes, with cooling, into a field of one-pyroxene crystallization. No data are available from volcanic rocks, but all data on plutonic rocks, of which the author is aware, have been assembled.

At least twelve pairs of co-existing pyroxenes from igneous rocks have been chemically analysed and appear in the literature (two are those used by Hess, 1941) and are plotted on Text-fig. 1. Ten of these are known to be pairs from the same rock. The other two (Stillwater)

are from rocks which are almost adjacent (in one case) or within the same general lithological horizon. The reason for including the latter two pairs will be clear from the diagram—they present points of intersection which lie at a *minimum* distance from the hypothetical $Wo_{75}$ position, and it is clear from their petrological associations that had each pair come from the same rock then they would have served only to further emphasize the deviation of the tie-lines (see below and p. 338).

The big layered basic intrusions provide the best evidence, amongst igneous rocks, of mineral assemblages developed by a very slow cooling process in which ample time was available for equilibrium to be reached between the separating crystal phases and the liquid. Care must be taken to distinguish carefully between the crystals precipitated from the main body of the magma to form a cumulus of crystals, and those formed from the relatively smaller amount of trapped intercumulus liquid ; the latter may in fact be in equilibrium with cumulus crystals from a higher level in the intrusion. However, poikilitic crystals may not necessarily be out of equilibrium with the cumulus crystals they enclose, for they may be of the heteradcumulate (unzoned) type. Such detailed diagnosis of the texture of a rock and the character of its constituent minerals, produced by the often complex processes associated with crystal accumulation, is essential if the information contained in these rocks is to be best used (see Wager, Brown, and Wadsworth, 1960, for a description of cumulate rocks and processes).

The Skaergaard intrusion contains a great thickness of rocks in which a calcium-rich and a calcium-poor pyroxene were forming together as a cumulus phase, and three of these pairs have been analysed. In addition, the earliest formed cumulate of the intrusion, the gabbro-picrite, contains an augite poikilitically enclosed by an orthopyroxene. As the latter is unzoned, it is unlikely that it is far from a composition in equilibrium with the enclosed augite. Thus analyses of four pyroxene pairs have been made for the Skaergaard, and published (Brown, 1957).

The Stillwater data were given by Hess in 1949, but recently in the Stillwater Memoir (Hess, 1960, pl. 2) it is possible to see the positions of the specimens within the intrusion and to use, with some reservation (see p. 338) specimens from different but neighbouring rocks. Three analysed pairs are available from the Stillwater, one being a pair from the same rock (EB41), one from adjacent rocks (EB43 and 130) and the third pair from rocks which are not very close but are both from just below the Pillow Troctolite of the Anorthosite Zone (EB38 and 175). As mentioned above, the latter two pairs are only used to give values of minimum deviation from the $Wo_{75}$ position.

The Bushveld data for two pairs were given by Hess in 1949 (and in 1960) as were data for a pair from the Goose Creek and from the Lambertville diabase.

Finally, a specimen not originally plotted for comparison with Skaergaard by Brown (1957, Text-fig. 3) is given here—the " porphyritic gabbro " of Kilauea which, from the published description (Muir and Tilley, 1957, pp. 243–4) may be classified as a normal igneous rock.

(*c*) *Discussion.*—The twenty-four pyroxene compositions plotted on Text-fig. 1 are all based on a chemical analysis of each mineral. Twenty of these (indicated by a circle) consist of co-existing pairs taken from ten igneous rocks. The other four (indicated by a square) are from four separate Stillwater rocks, and the two tie-lines join pairs from neighbouring rocks. The error in these two tie-lines cannot be great, when the distance apart is considered of the layers from which each member of a pair was collected, but it should be recognized. It does not weaken the argument in this paper, for the intersection points for these two pairs are closer to the $Wo_{75}$ position than they would be for pyroxene pairs from identical rocks.

For discussion, it is convenient to list the values (see Table 1) for the position at which each projected tie-line intersects the Ca–Mg join in Text-fig. 1. Wilson (1960) called these intersection points XMg (i.e., on the Ca–Mg join) and XFe (on the Ca–Fe join), and these useful abbreviations are used here.

The notes (Max. and Min.) in Table 1 can be explained briefly. In the case of No. 1, the orthopyroxene is intercumulus so that, if the considerations mentioned on p. 337 do not apply, the orthopyroxene which would have been in equilibrium with the cumulus augite of this rock would be more magnesian than the analysed specimen, swing the tie-line clockwise, and give a lower XMg value. For No. 8, the augite may be intercumulus (cf. Hess, 1949, p. 645), so that the augite which would have been in equilibrium with the cumulus bronzite in this rock would be more magnesian than the analysed specimen, swing the tie-line anti-clockwise, and give a higher XMg value. In No. 5, the orthopyroxene came from a higher rock than the augite, so the tie-line should be turned clockwise, and give a higher XFe value. In No. 6, the orthopyroxene also came from a higher rock than the augite, the tie-line should also be turned clockwise, and give a lower XMg value. Apart from No. 8, these pairs indicate a minimum degree of divergence from the point X Mg = 25.

The values for X Mg on Table 1 represent all which are apparently available for igneous pyroxenes. It is a matter for personal choice whether or not the petrologist is prepared to say that they do not indicate appreciable divergence from the long-accepted figure for XMg of about 25. What is more important, however, is to ask *why* the figure of XMg = 25 should be accepted. There is no evidence for it, anywhere in the literature, from chemical information. Not one tie-line in Text-fig. 1 intersects the Ca–Mg line exactly at XMg = 25·0. The

deviations from this value are impressive and, in the author's opinion, invite only one conclusion. That is, that in co-existing igneous pyroxene pairs the calcium-poor pyroxene is almost always enriched in Fe/Fe+Mg relative to the calcium-rich pyroxene, causing the tie-line to project to the left of the Ca-apex of the triangular diagram. However, although the degree of relative enrichment may follow

TABLE 1.—INTERSECTION VALUES FOR ANALYSED IGNEOUS PYROXENE PAIRS.

| Locality | | Rock No. | Ca | Mg | Fe | XMg | XFe | Note |
|---|---|---|---|---|---|---|---|---|
| Skaergaard | 1 | EG 4526 | 42·4 | 47·9 | 9·7 | 15·5 | — | Max. |
| | | | 3·9 | 77·6 | 18·5 | | | |
| | 2 | EG 4385A | 37·7 | 41·4 | 20·9 | 21·0 | — | |
| | | | 9·0 | 55·8 | 35·2 | | | |
| | 3 | EG 4341 | 35·8 | 37·8 | 26·4 | 20·5 | — | |
| | | | 8·9 | 48·2 | 42·9 | | | |
| | 4 | EG 4430 | 34·6 | 37·0 | 28·4 | 23·0 | — | |
| | | | 8·6 | 45·5 | 45·9 | | | |
| Stillwater | 5 | EB 175 | 42·3 | 46·0 | 11·7 | — | 5·0 | Min. |
| | | EB 38 | 3·7 | 79·7 | 16·6 | | | |
| | 6 | EB 43 | 41.0 | 45·0 | 14·0 | 3·5 | — | Max. |
| | | EB 130 | 3·6 | 73·1 | 23·3 | | | |
| | 7 | EB 41 | 39·5 | 42·5 | 18·0 | 22·0 | — | |
| | | | 10·0 | 58·0 | 32·0 | | | |
| Bushveld | 8 | 7666 | 45·0 | 50·0 | 5·0 | 9·0 | — | Min. |
| | | | 2·8 | 87·5 | 9·7 | | | |
| | 9 | 7493 | 37·6 | 41·1 | 21·3 | 14·0 | — | |
| | | | 8·5 | 57·0 | 34·5 | | | |
| Lambertville | 10 | LAM 12 | 37·0 | 39·7 | 23·3 | 34·5 | — | |
| | | | 9·3 | 44·4 | 46·3 | | | |
| Goose Creek | 11 | TR-1 | 36·7 | 38·4 | 24·9 | 17·0 | — | |
| | | | 9·5 | 50·8 | 39·7 | | | |
| Kilauea | 12 | — | 39·5 | 49·8 | 10·7 | 26·0 | — | |
| | | | 5·3 | 73·4 | 21·3 | | | |

Compositional data from the following sources :—1–4 (Brown, 1957, Table 1) ; 5–9 (Hess, 1960, Tables 3, 4, and 9) ; 10–11 (Hess, 1949, Analyses 32, 33, 41, and 42) ; 12 (Muir and Tilley, 1957, Table 4).

some law, it cannot be easily explained by suggesting that all the tie-lines meet at a single point. If an average position for their intersection of the Ca–Mg line is desired, then it is to be found by averaging the results in Table 1. The figure thus obtained (treating XFe as a negative figure) is XMg = 18 (approx.).

## TIE-LINES FOR PYROXENES FROM ULTRABASIC INCLUSIONS IN BASALTS

Ross, Foster, and Myers (1954) published a wealth of chemical data which included the compositions of nine co-existing pairs of calcium-rich and calcium-poor pyroxenes. They were not included in the previous section because some doubt may exist as to whether the rocks from which they were separated can be classed as igneous, in the sense that the minerals have crystallized from a cooling magma with which

equilibrium was maintained, or whether they have been since metamorphosed to some degree. Eight of the pyroxene pairs are from ultrabasic nodules included in basalts, and their unusually high $Al_2O_3$ content suggests crystallization or re-crystallization under a fairly high pressure. The ninth pair are from the type websterite of North Carolina, and they contain amounts of alumina similar to the higher level igneous pyroxenes discussed in the previous section. However, the textures of some of the nodules, as well as their isotopic composition, suggest an origin similar to the igneous process of crystal accumulation from a basaltic liquid (see Wager, 1958, p. 7). If this is so, then the tie-lines of their co-existing pyroxenes ought to be considered in relation to those shown in Text-fig. 1.

When the tie-lines were constructed, the following values were obtained for their intersection of the Ca–Mg and Ca–Fe sides of the ternary diagram. The numbers refer to the specimen numbers given by Ross *et al.* (1954, Tables 5 and 6). No. 1, XFe = 3·5 ; 2, XFe = 2·0 ; 3, XFe = 3·0 ; 4, XFe = 1·5 ; 5, XFe = 3·0 ; 7, XMg = 3·0 ; 8, XFe = 0·5 ; 9, XFe = 7·0 ; 13 (websterite), XMg = 31·0. Thus, excluding the last figure, the intersection point for the pyroxene pairs from basaltic nodules averages at XFe = 2·0.

This result is so different from that obtained for the igneous pyroxenes, listed in Table 1, that it may well be significant. Certainly, considering the evidence presented by Ross et al., it would suggest that they are not entirely justified in claiming (op. cit. p. 732) that the nodules and the dunites (which include the websterite) are derived from the same source, and have had similar genetic histories. The websterite stands isolated from the nodules, with respect to its pyroxene tie-line, and close to the igneous group. Significantly, the $Al_2O_3$ content of the websterite and dunite pyroxenes is not high, but is similar to that of igneous pyroxenes, and it may be this factor which is primarily connected with the different Fe:Mg distribution between the co-existing pyroxene pairs. If, on the other hand, other evidence is in favour of these nodules being " igneous ", then the tie-line data provide even stronger evidence against the acceptance of a standard value of XMg = 25 for the igneous pyroxene tie-line intersection.

## Tie-lines for Pyroxenes from Metamorphic Rocks

The compositions of co-existing pyroxene pairs from metamorphic rocks have been discussed by Howie (1955 and 1958), Muir and Tilley (1957 and 1958), and Wilson (1960). Such studies are not in the province of the present paper, but the observations made by Wilson are such as to provocate comment, particularly in the light of what has been said in the previous sections about igneous pyroxenes.

Wilson (1960) measured the optical properties of pyroxene pairs from certain metamorphic rocks (chiefly charnockitic granulites and charnockitic granites from Central Australia), and combined these with published chemical and optical data on pyroxene assemblages from other metamorphic rocks or rock suites. The tie-lines for about fifty metamorphic pairs were thus obtained, together with the positions at which their projections intersected the sides of the Ca–Mg–Fe triangle. From this information Wilson discusses at length the manner in which the metamorphic tie-line intersections differ from the igneous tie-line intersection (at $Wo_{7\cdot5}$), and its implication, and further sub-divides the metamorphic rocks (on optics alone) according to the position of their intersection.

The fact that Wilson accepts without question the figure of $Wo_{7\cdot5}$ for igneous pyroxenes, suggested by Hess twenty years ago, and disregards all subsequent data, is surprising. This is particularly so because this figure is taken by him to be " normal " and indicative of the ideal intersection point for the tie-lines of pyroxenes in equilibrium ; any deviation from this point is considered " anomalous " and indicative of departure from equilibrium conditions (e.g., op. cit. p. 14). Thus the $Wo_{7\cdot5}$ position forms the basis of Wilson's paper, even though it is unacceptable through work done subsequent to 1941.

As the igneous pyroxene pairs have been plotted in the present paper only if chemically analysed, a direct comparison with metamorphic pyroxenes can best be made, first, by considering only chemically analysed metamorphic pyroxene pairs. Fifteen such pairs can be found in the literature, and Wilson (1960, Table 5) gives the XMg values for fourteen of them. The data for the fifteenth pair are given by Howie (1958, Table 2), from a basic charnockite from the Lafit Mountains, Sudan, and the XMg value is 23·0. The XMg values show a great spread, comparable with those for the igneous pyroxenes, and range from 3·0 to 27·5 (e.g. chemical column of Table 2). They average, taken with Howie's pair, at exactly 18·0. The igneous pyroxenes, as seen from Table 1, show a spread from − 5·0 to 34·5. They average, taken with the websterite and amounting to thirteen pairs, at exactly 18·0.

From these results the following points emerge : (1) The intersection value for the igneous pyroxenes is not at XMg = 25, but averages at XMg = 18. (2) The intersection value for metamorphic pyroxenes also averages at XMg = 18. (3) The spread of values for both igneous and metamorphic assemblages are great, and similar. Thus it is not possible to draw any conclusions as to the degree of metamorphism and the state of equilibrium of the metamorphic assemblages by comparison with the igneous assemblages. In view of the identical average values, it may seem tempting to suggest that the igneous and

metamorphic pyroxene assemblage tie-lines intersect at the same point. However, in view of the great spread of values in both groups, the present author would be less commital and agree with Muir and Tilley (1958, p. 406) that " in metamorphic assemblages carrying two pyroxenes the tie-line trends show no significant departure from that recorded from plutonic igneous assemblages ". It is further hoped that, in view of the data presented, no special significance will be attached to the value of XMg = 18, even remotely comparable with the significance which has, in the past twenty years, been attached to the value of XMg = 25.

TABLE 2.—DISCREPANCIES BETWEEN INTERSECTION VALUES OBTAINED BY CHEMICAL AND OPTICAL DETERMINATIONS ON METAMORPHIC PYROXENES

| Specimen | XMg Values (XFe expressed as negative) | | |
|---|---|---|---|
| | Chemical | Optical | Discrepancy |
| 1 | 25·5 | 8 | 17·5 |
| 2 | 8 | −16 | 24 |
| 3 | 23 | 1 | 22 |
| 4 | 3 | −15 | 18 |
| 5 | 23 | −6 | 29 |
| 6 | 20·5 | 15 | 5·5 |
| 7 | 22 | 32·5 | 10·5 |
| 8 | 14·5 | 5·5 | 9 |
| 9 | 15 | 11 | 4 |
| 13 | 22·5 | 8·5 | 14 |
| 48 | 5 | −16 | 21 |
| 49 | 21·5 | 14 | 7·5 |
| 51 | 16 | 19 | 3 |
| | | | Average = 14 |

The figures in all but the last column are taken directly from Wilson, 1960, Tables 4 and 5.

## TIE-LINES FROM OPTICAL MEASUREMENTS

Despite the wide spread of XMg values obtained from the previous consideration of fifty-six chemically analysed pyroxenes, it may be that more chemical data will allow a pattern to emerge (cf. the additional data on sixteen analysed pyroxenes from basaltic nodules, p. 340). On the other hand, there seems little to be gained in this direction from making a series of optical determinations (advocated by Wilson, 1960) in order to learn something, from the pyroxene tie-lines, about the rock's history. The reason for this is that a pattern has not yet emerged from the chemical data, and an attempt to establish one by indirect optical means can only be advantageous if the correlation between optical and chemical data are dependable. This correlation

can be shown to be suspect by reference to the previous literature (see p. 335), but a more illuminating example can be taken directly from the data presented, in support of a different argument, by Wilson. Twenty-six of the chemically analysed metamorphic pyroxenes have each been subjected to detailed optical study (Wilson, 1960, Table 4), and the XMg values obtained by chemical means can, therefore, be compared with those obtained from optics (see Table 2). The optical data presented are obtained from $\gamma$ for both pyroxenes, in view of the reservations made by Wilson (op. cit. p. 12).

The discrepancies between values obtained by the two methods are so great that further comment, as to the value of optical methods in obtaining a precise XMg value, seems unnecessary.

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# CORRESPONDENCE

## CO-EXISTING PYROXENES

SIR,—Three papers dealing with the distribution of Mg and Fe between co-existing orthopyroxene and clinopyroxene have recently appeared in the *Geological Magazine*. Muir and Tilley (1958) concluded that the distribution is the same for metamorphic and plutonic igneous rocks. Wilson (1960) suggested that the distribution may vary as a function of temperature. O'Hara (1960) arrived at a conclusion which concurs with that of Muir and Tilley. In each paper the distribution is discussed in terms of tie-lines in a three-component diagram; the intersection point of a side of the diagram and the projection of a tie-line is taken as a measure of the distribution.

The subject can be examined from a thermodynamic viewpoint (Ramberg and DeVore, 1951). The distribution of Mg and Fe between two minerals can be expressed by use of a distribution diagram (Kretz, 1959). When this is done for co-existing pyroxenes from six Indian charnockites studied by Howie (1955) we obtain a curve, the shape of which indicates that both phases are ideal mixtures (Mueller, 1960). Thus the distribution coefficient can be taken as a direct measure of the distribution of Mg and Fe between the two minerals. The distribution coefficient ($K_D$) has the form (Ramberg and DeVore, 1951):

$$K_D = \frac{X^o}{1 - X^o} \cdot \frac{1 - X^c}{X^c}$$

where $X^o = Mg/(Mg + Fe^{2+})$ in orthopyroxene, and $X^c = Mg/(Mg + Fe^{2+})$ in clinopyroxene.

Values of $K_D$ for orthopyroxene-clinopyroxene combinations from two groups of rocks are presented in Table 1. The first group of rocks was supposedly equilibrated at metamorphic temperatures and the second group at near-liquidus temperatures. Note that $K_D$ is nearly constant in the first group and consistently higher in the second group.

TABLE 1

| *Spec. No.* | $K_D$ | *Source* |
|---|---|---|
| Metamorphic rocks: | | |
| 3709 | 0·553 | Howie (1955) |
| 4645 | 0·510 | " " |
| 2270 | 0·532 | " " |
| 2941 | 0·561 | " " |
| 4642A | 0·567 | " " |
| 115 | 0·598 | " " |
| O | 0·578 | Muir and Tilley (1958) |
| S | 0·564 | " " " " |
| 35–5 | 0·541 | Ward (1957) |
| 35–6 | 0·548 | " " |
| 35–8 | 0·534 | " " |
| 35–9 | 0·515 | " " |
| 35–19 | 0·525 | " " |
| Igneous rocks: | | |
| 1 | 0·716 | Brown (1957) |
| 1.H | 0·690 | Muir and Tilley (1957) |
| A | 0·710 | Carstens (1958) |
| B | 0·762 | " " |

Three rocks studied by Muir and Tilley (1958) are of special interest. The specimens are numbered X, R, and T, and contain mineral pairs which yield $K_D$ values of 0·856, 0·776, and 0·782 respectively. These are metamorphic

rocks with values of $K_D$ characteristic of igneous rocks. Specimen X is a metamorphosed basalt and the remaining two may also have experienced igneous temperatures, as indicated by the presence of olivine. The possibility exists that $K_D$ in the three rocks is a relic of a pre-metamorphic high-temperature equilibrium.

On theoretical grounds we find that $K_D$ is dependent on (1) temperature, (2) pressure (Ramberg and DeVore 1951), and (3) variable concentrations of a third element (Kretz, 1961). It is reasonable to suspect that large variations in $K_D$ (Table 1) have resulted from a variation in the equilibration temperature. It is uncertain, however if small variations in $K_D$ (as found in the first group of rocks in Table 1) may be attributed entirely to the temperature dependence of the distribution coefficient.

The subject will be discussed in detail in a forthcoming paper. At present I should like to make a special plea to abandon the meaningless operation of projecting tie-lines and to propose that we discuss the distribution of Mg and Fe in co-existing pyroxenes in terms of the distribution coefficient.

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Ottawa,
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*25th January, 1961.*

## CO-EXISTING PYROXENES IN IGNEOUS AND METAMORPHIC ROCKS

SIR,—In recent issues of this magazine several papers have been published, dealing with co-existing pyroxenes in igneous and metamorphic rocks. The authors of these papers were concerned with the possibility of using the pyroxene tie-lines as a geological thermometer. It may be useful to consider the theoretical aspects of this problem.

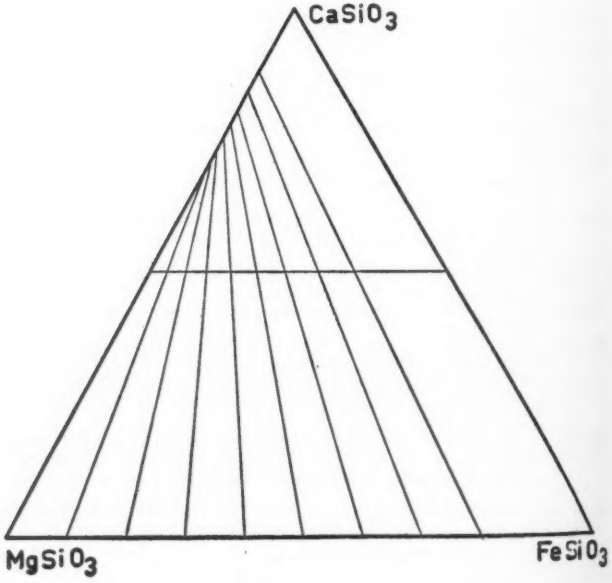
The chemical equilibrium between calcium-rich and calcium-poor pyroxenes may be written as follows :

$$CaFeSi_2O_6 + MgSiO_3 \rightleftharpoons CaMgSi_2O_6 + FeSiO_3$$

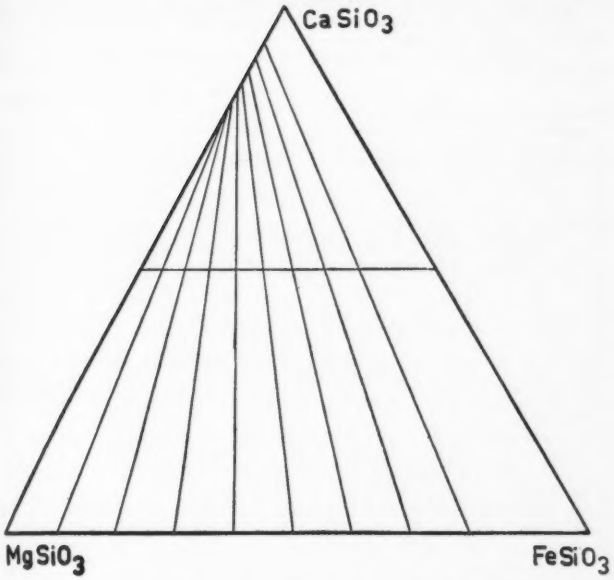
the condition of equilibrium being if the minerals are ideal solutions

$$\left(\frac{Fe^{++}}{Mg}\right)_{Ca\text{-poor}} = \left(\frac{Fe^{++}}{Mg}\right)_{Ca\text{-rich}} \times K(T) \quad . \quad . \quad (1)$$

Using the chemical analyses of co-existing pyroxenes found in the literature, one finds that the K value calculated from equation (1) is on the average 1·8 for metamorphic rocks (Mueller, 1960), 1·4 for igneous rocks, and 1·2 for olivine nodules. Since K is known to within a standard relative deviation of about 10 per cent the conclusion is reached that the pyroxene associations in rocks constitute a rather insensitive geological thermometer.




TEXT-FIG. 1.—The pyroxene equilibrium diagram based on the hypothesis that equation (1) is verified. Isothermal section drawn for K = 1·8 and corresponding approximately to metamorphic conditions.




TEXT-FIG. 2.—Isothermal section drawn for K = 1·4 and corresponding approximately to igneous conditions.

Equation (1), which is approximately obeyed in natural systems, is represented graphically in Text-figs. 1 and 2. These diagrams, corresponding to K-values of 1·8 and 1·4 respectively, show what little difference should be expected between igneous and metamorphic tie-lines. It should be noted that, in Text-figs. 1 and 2, tie-lines are drawn between two points, one of which lies on the enstatite-ferrosilite join, and the other on the diopside-hedenbergite join ; in other words, consideration is given only to the iron-magnesium ratio in the minerals involved, the calcium content being disregarded.

Another conclusion emerges from an examination of Text-figs. 1 and 2 : the point where a tie-line intersects the left-hand side of the triangle cannot be expected to be a measure of temperature. Actually, tie-lines corresponding to the same temperature, i.e. belonging to the same isothermal section of the equilibrium diagram, do not meet at a point on the left-hand side of the triangle ; they intersect outside the triangle. Such a conclusion may seem to contradict the facts observed since, in diagrams representing natural occurrences, tie-lines sometimes intersect within the triangle. The contradiction is, however, only apparent. Diagrams representing natural occurrences cannot be directly compared with Text-figs. 1 and 2 because in such diagrams (1) tie-lines in the magnesium-rich region correspond to a temperature somewhat higher than tie-lines in the iron-rich region ; (2) mineral compositions are plotted with due regard to the calcium content. In addition it should be observed that the upper part of a tie-line is very sensitive to any error in the determined composition of the calcium-rich pyroxene.

May I finally add that, in the study of mineralogical equilibria, the practice of adding ferric iron and manganese to ferrous iron has no justification and can only bring confusion?

I am indebted to Dr. G. M. Brown for a stimulating discussion on pyroxene tie-lines.

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13*th February*, 1961.

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## TERTIARY OF BARBADOS, W. I.

SIR,—In a recent paper on exploration results in Barbados, P. H. Baadsgaard (1960) introduced several insufficiently defined stratigraphic terms that appear to offer no improvement on the definitions of A. Senn (1940, 1948).

*Scotland Formation*. Senn's Lower Scotland Formation includes the Morgan Lewis and the Walkers beds. Baadsgaard places his " P-unit " below the Walkers beds, and suggests the term " *River Formation* " for the three members. He describes his " P-unit " as " a new subsurface unit of black, sometimes fossiliferous silty shales and mudstones with few thin fine sandstones ". Senn (1940, p. 1553) gave a thickness of 240 metres for the exposed type-section of the Walkers beds. In a bore-hole 650 metres of these beds were penetrated without having reached the bottom. The " P-unit " appears to form part of the lower, unexposed Walkers beds, hence the term " River Formation " appears to be unnecessary. If Baadsgaard wants to introduce a new subdivision he should bring out the lithological and palaeontological differences, and discuss the boundaries as seen in the Friendship well.

Senn's Upper Scotland Formation consists of Mount All, Chalky Mount, and Murphys beds. The " *Bruce Vale Formation* " of Baadsgaard embraces the same three members, and for this reason appears to be synonymous with Senn's Upper Scotland Formation.

Baadsgaard's " *T-unit* " is " a new subsurface stratigraphic element beneath the superficial Oceanic section in Friendship–1, Springvale–1, and Ince–1 ". This unit " consists of blue or purple claystones with thin siltstones and, locally, bedded sandstones and septarian limestones ". In Text-fig. 2 the " T-unit " crosses the time lines from Middle Eocene to Upper Oligocene analogous to the older " P-unit ". The " T-unit " of bore-hole Ince–1 is covered by beds of Baadsgaard's " Oceanic Group ", for which no age is given. Baadsgaard shows in his Text-fig. 3 at least " three superimposed major structural plates " which he assumes to be " sheet-like displacements with extensive overturned sections ". In none of the seven exploration wells is the " T-unit " involved in these tectonic complexities. It always rests on top of Senn's Scotland Formation, and below Baadsgaard's " Oceanic Group ". Where palaeontologically identified in Ince–1, the " T-unit ", Bruce Vale and River formations yielded faunas compared by Bronnimann with the Herrera member of the Cipero Formation of Trinidad. This makes them not older than Aquitanian and not " good Oligocene " as stated by Baardsgaard. Oceanic type marls are known from lower horizons of the Miocene, from Oligocene and from Eocene below which follows the Scotland Formation. Unless Baadsgaard can give strong proof of the highly diachronous nature of all these units it seems wiser to consider the younger, Miocene parts of the

Bruce Vale and River formations, together with the " T-unit ", as a flysch facies within younger formations and not as part of the Scotland Formation. An analogy could be drawn with Trinidad where the muddy flysch facies of the Nariva Formation interfingers with the Cipero marls.

Baadsgaard's term " *Scotland Group* " is premature as long as there are no well-defined formations to compose such a group.

*Joes River Formation.* Baadsgaard believes the Joes River Formation to be a kind of shear-zone breccia locally mobilized by gas, oil, and water, and occurring at different levels in his " Scotland Group ". He is, therefore, not prepared to give the Joes River beds the rank of a formation. Senn considered the Joes River Formation to be the product of mud volcanoes, partly interbedded with normally deposited silts. To Hess (1937, p. 73 and 1938, p. 89) these rocks appeared " to be of the nature of a submarine landslide " which took place during the downward movement of the West Indian tectogene. The Joes River Formation of Senn is stratigraphically defined. Though shear zones may have given rise to the formation of " pebbly silts ", and are possibly responsible for the presence of large blocks of Paleocene *Discocyclina* limestone and other rocks, we must look upon the pebbly silt as representing mudflows separated one from the other by layers of normally bedded silt and clay. Senn (1940, p. 1576) considered the massive silt breccia and boulder beds to be resting with apparent unconformity on the Scotland beds. Baadsgaard found rocks of a lithology comparable to that of the Joes River Formation in Springvale–1 from 2,000 to 9,650 feet in the highly disturbed Scotland Formation. Senn (1940, p. 1574) already considered the Joes River pebbly silt to have been derived from various parts of the Scotland Formation, and to have ascended in vents that could now be interpreted as " mobilized shear zones ". The product of these shear zones could have been extruded, spread, and normally deposited as now found in the basal part of the Oceanic Formation. The pebbly silt is incompetent. Under orogenic pressure it was even injected into cracks of the Oceanic marls. Drilling in Trinidad has proved that Miocene pebbly clays with large blocks and slip-masses of pre-Miocene rocks form diapiroid bodies comparable to salt plugs. The writer saw no difference between the Joes River Formation and his " marine-slope clay-breccias " (Kugler, 1953, p. 51) which he described from Trinidad. Marchant (1960), who described the Ancon " Clay Pebble-beds " of Ecuador, also studied those of Trinidad. He found both to be very similar sedimentary deposits, formed by submarine sliding. Baadsgaard (p. 24) states " . . . transitions can be traced from structureless Joes River to normal country rock " ; such transitions appear to exclude the presence of mobilized shear zones.

*Oceanic Formation.* Already Senn, and later Beckmann, have shown that the Oceanic Formation represents an uninterrupted sequence of foraminiferal marls with planktonic assemblages ranging in age from the middle Eocene *Porticulasphaera mexicana* zone to the top of the Codrington College marl, which belongs to the base of the *Catapsydrax dissimilis* zone of the Cipero Formation of Trinidad (Beckmann, 1956, p. 131). It is difficult to conceive how this uniformly widespread formation should be replaced by the Scotland Formation, or a facies of it, as is shown in Baadsgaard figure 2. The results of the bore-holes do not support his interpretation. No interdigitation of Oceanic type sediments with Scotland type sediments is known from surface exposures.

*Bissex Hill Formation.* This formation with its shallow water fauna overlaps the Oceanic Formation at Bissex Hill. It belongs to the *Globigerinatella insueta* zone and, in facies and age, is to be correlated with the Ste. Croix beds of Trinidad (Beckmann, 1956, p. 130). The Ste. Croix limestone is a member of the Cipero Formation with its calcareous clays and marls. In Trinidad deeper water Cipero marls were laid down contemporaneously with the Ste. Croix member, while in Barbados the lower part of the " Upper Oligocene *Globigerina* Marl " of Senn seems to be an age equivalent of the Bissex Hill Formation (according to Beckmann, 1956, p. 140).

*Upper Oligocene Globigerina Marl.* Senn (1948, p. 214) described this marl in detail. He knew it only from water wells, but Beckmann (1956, p. 140) states : " The Upper *Globigerina* marls of Senn are now known from surface outcrops (e.g., at St. Mark's near Conset Bay). They represent the *Globorotalia fohsi* and *Globigerinatella insueta* zones. Therefore, they are not older than the Bissex Hill Formation, as Senn (1948) assumed, but younger ." Finally Beckmann states : " The sedimentary facies of the ' Upper Oligocene *Globigerina* marls ' is closely related to that of the Codrington College member of the Oceanic formation." Baadsgaard gives no description of his " *Waterwell Marls* ". There is little doubt, however, that they are the same as Senn's " Upper Oligocene *Globigerina* Marl ". It is here suggested that this marl be given the name " *Conset Marl* ". Senn (1948, p. 215) believed that there are indications of an unconformity between his Upper Oligocene *Globigerina* Marl and his Oceanic Formation. According to Baadsgaard, exploration-well Adams Castle–1 penetrated 5093 feet of " an unbroken sequence from upper Eocene to upper Oligocene ". Apparently no " T-unit " or Bissex Hill Formation ws observed. The bore-hole proved the presence of a deep, post-middle Eocene trough that existed in the S.W. corner of Barbados. Should detailed palaeontological studies of the bore-hole samples corroborate Baadsgaard's claim, one would be justified in extending the term Oceanic Formation to include also marls of Miocene age. In such case one could treat the Bissex Hill Formation and the Conset Marl as members of the Oceanic Formation.

*Acknowledgment.* I am obliged to Mr. J. B. Saunders of Texaco Trinidad Inc., for contributions to this note.

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*22nd April, 1961.*

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# REVIEWS

MORPHOLOGY, CLASSIFICATION, AND LIFE HABITS OF THE PRODUCTOIDEA (BRACHIOPODA). By HELEN MUIR-WOOD and G. ARTHUR COOPER. Pp. xi + 447, with 8 figures and 135 plates. Geological Society of America, Memoir 81. 1960.

So much systematic palaeontology appears nowadays in small piecemeal contributions that a comprehensive revision of a major group is always welcome. This work shows clearly the amazing diversity that developed within the Productoidea, and it will be valuable to all concerned with the study of Upper Palaeozoic faunas. Some readers, noting that new taxa include ten out of nineteen families, thirty-one out of forty subfamilies, and sixty-six out of 167 genera, may conclude that the modern zeal for " splitting " has reached new heights in brachiopod systematics. But this would not be fully justified, for this work describes many genuinely new forms—or forms newly known in great detail—from the exquisitely well-preserved Permian faunas of Texas. The contrast between these and less well-preserved faunas accounts for the work's chief drawback : many species are not assigned here to any genus, because the material on which they were based is not well enough preserved to show the relevant diagnostic features. But since it would be inexcusable to ignore the information that the silicified faunas provide, this dilemma seems unavoidable.

Introductory sections deal with techniques of study, morphology, life habits, origins of the group and principles of classification. The section on morphology is full and detailed, and defines many new features which figure later in the generic descriptions. It is, therefore, rather surprising to find, in the diagnoses of some of the higher taxa, the survival of such non-morphological and question-begging terms as " specialized ", " degenerate," and " primitive ". The section on life habits provides a useful summary of the ways in which the shells were related to the substratum, though it adds little to previous interpretations. The origin of the productoids is left uncertain, and a polyphyletic origin is not ruled out. Chonetoids are regarded as only distantly related. Two major lines—the Strophalosiacea, cemented by beak or spines throughout life, and the Productacea, rarely cemented after early life—are thought to have been distinct almost from the beginning. The bulk of the volume consists of the generic descriptions, which are clear, well arranged, and illustrated by excellent collotype plates.

M. J. S. R.

GEOLOGICAL NOMENCLATURE. Edited by A. A. G. SCHIEFERDECKER. pp. xvi + 523. J. Noorduijn en Zoon N.V., Gorinchem, Holland, 1959. Price 45 guilders.

This successor of Professor Rutten's work in 1929 is another triumph for Dutch geology and language. Here is a prodigious amassing of 5,489 groups of terms. They are arranged systematically according to subject matter and similarity of idea. Each group is headed by the English term, with such synonyms and explanation as seem suitable, followed in turn by the Dutch, French, and German equivalents. If perhaps Spanish and Russian could be added later a still greater service would be done. The numbered groups are then followed by a union index in which words in the four languages are arranged alphabetically and refer to the appropriate numbers of the glossary sections. One thus has easy access to a vast collection of terms, the arrangement of which has benefited by the labour of dozens of specialists.

Unfortunately few of the experts could have been sufficiently at home in the English language to decide some of the nicer points of translation. Thus, perhaps inevitably German, French, and Dutch words are forced into English form. *Parallelization* is given as an alternative English form for correlation,

and *stereogram* for block diagram. It cannot be denied that such terms are used extensively by many foreign writers in English and no doubt taken from these sources in due course by English readers who cannot but be grateful that so many take the trouble to write in English. English has become a universal language for science, and if the centre of gravity in geological terminology has moved far across the Atlantic it is only fair to accept the contributions that other communities can and will make to it.

However, there is a limit to indiscriminate multiplication of jargon and many uses could have been subjected to more critical scrutiny. Our old friends *pitch* and *plunge* for instance are not distinguished and there are many other instances where the glossary could have led to a refinement rather than a coarsening in the precision of language—but over so wide a field it is ungenerous to cavil. Moreover, a lexicon which sets out to be authoritative would still not survive criticism and if the choice is between quality and quantity the reviewer prefers the latter in this case since it is usually better to find some indication of a word than not to find it at all (which is the usual experience)—and it still allows the user his own discretion.

Quite apart from their inherent use for reference, however, dictionaries and glossaries, like directories and lists of births, marriages and deaths, appeal to a significant section of the community as reading or at least browsing matter. The work under consideration will come high on their list, since the arrangement, like Roget's *Thesaurus*, depends on association of ideas, and the appeal of a suggestive trail of thought leads the reader on. Such gems as *ombrogenous* (an adjectival term indicating that peat formation is a result of rainfall) add spice to the chase and yield rich trophies to the Lifeman—but the reviewer will desist from further exhibition of his ignorance.

W. B. H.

THE ROTATION OF THE EARTH. A GEOPHYSICAL DISCUSSION. By W. H. MUNK and G. J. F. MACDONALD. Cambridge Univ. Press, 1960. pp. xix (including Notation) + 323, with figs. and tables. Price 70*s*.

This book is concerned with changes in the position of the pole and in the earth's angular velocity. Some of these changes, such as the " Chandler wobble " with a period of 436 days, have been known for a long time ; some, such as the changes in the length of the day, have been discovered in the last few years and some, such as " polar wandering " may not exist at all.

The literature of the subject is so confused by errors in dynamics and in the treatment of observations that it is hard for the casual reader to follow it. The present authors have produced a most valuable treatment of the whole subject which makes it clear just where others have gone wrong and what may and what must be believed.

The book is mathematical and cannot be recommended as light reading for the tired geologist at the end of a day's field work. It is, however, enlivened by accounts of the history of the subject, including a discussion of a statement made by Senator Kefauver during the 1956 Presidential election which had been exposed as an error seventy years before by Jules Verne.

E.C.B.

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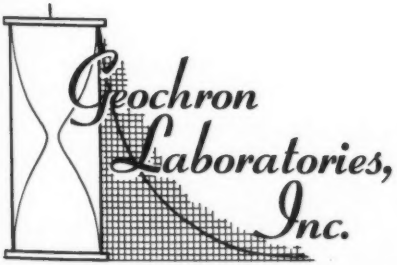


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